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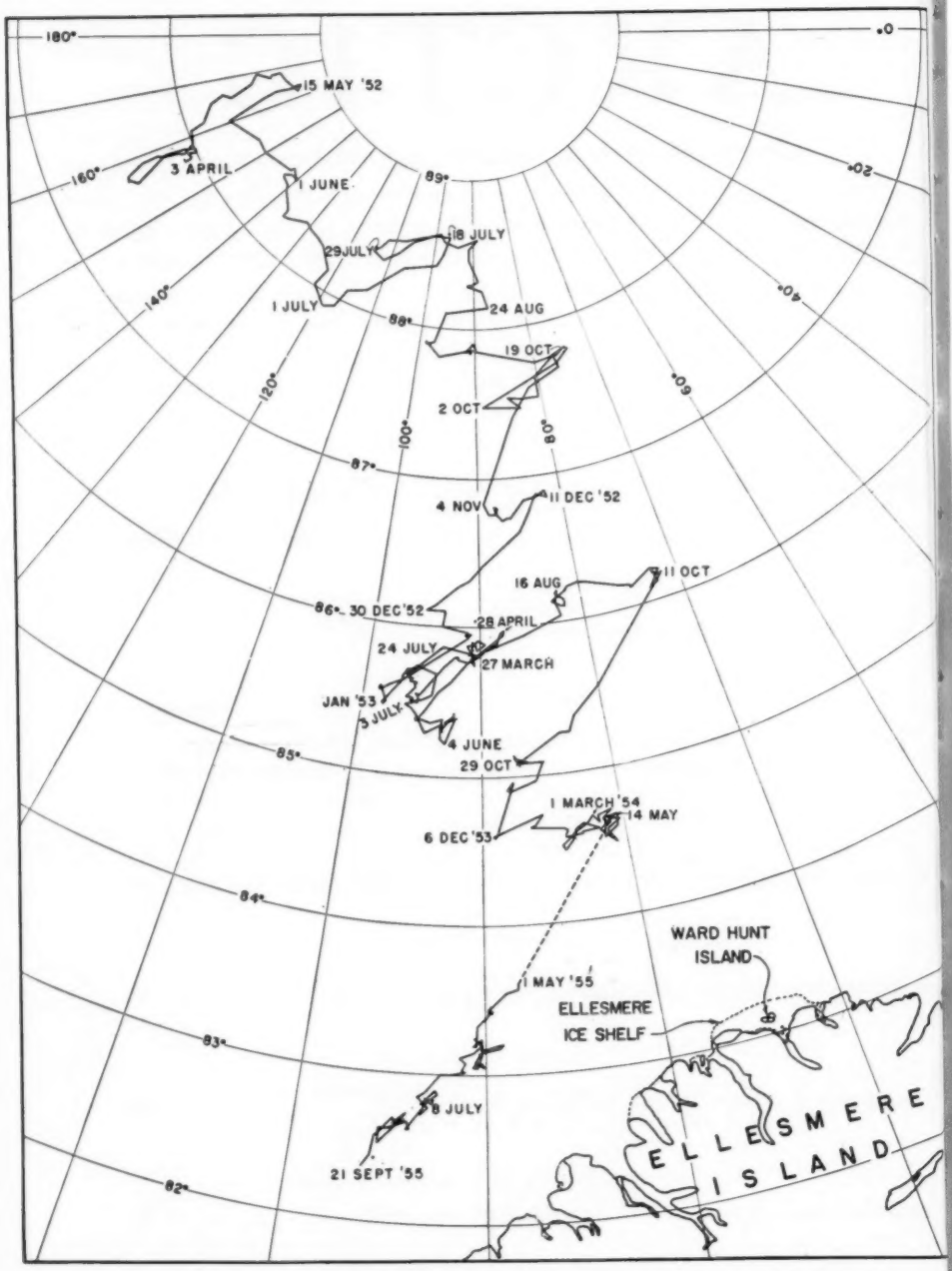


Fig. 1. Path of ice island T-3 during its period of occupancy shown in relation to the north coast of Ellesmere Island.

ARCTIC ICE ISLAND AND ICE SHELF STUDIES

Part I

A. P. Crary*

Introduction

In March 1952, in what was termed Project ICICLE, under the leadership of Lt. Col. Joseph O. Fletcher, the Alaskan Air Command of the United States Air Force landed a small party of men about 150 miles from the North Pole on T-3, one of three floating ice islands whose movements in the Arctic Ocean had been followed since the first sighting in 1946. Soon after the initial landings meteorological upper air observations were started by the Air Weather Service, and geophysical studies were begun by the Air Force Cambridge Research Center. The upper air weather station continued in operation for a total of 22 months until in 1954 at 84°40′N., 81°W. operations were abandoned. During this period of occupation various scientific studies were carried out periodically by personnel from the Air Force Cambridge Research Center, from the University of Southern California under contract to the Arctic Aeromedical Laboratory, from Woods Hole Oceanographic Institute under contract to the Office of Naval Research, and from SIPRE (Snow, Ice, and Permafrost Research Establishment). A small group of scientists reoccupied the island from April to September 1955 when it was in the general area of 83°N., 90°W.

In an attempt to correlate the character of the ice island with that of the Ellesmere Ice Shelf, the presumed source of T-3, two expeditions led by Geoffrey Hattersley-Smith of the Defence Research Board of Canada explored the northern shores of Ellesmere Island in the summers of 1953 and 1954 from the weather station Alert to Lands Lokk, with the main efforts being concentrated near Ward Hunt Island in 1954. Each expedition was accompanied by a geologist of the Geological Survey of Canada, and in the second expedition United States personnel from the Air Force Cambridge Research Center and from SIPRE also participated. Fig. 1 shows the track of T-3 during its period of occupancy and also shows the location of the northern Ellesmere Ice Shelf. In May 1952 a small party from T-3 briefly visited T-1, one of the largest of the known floating ice islands, when it was about 40 miles offshore from Ellesmere Island.

* Geophysics Research Directorate, Air Force Cambridge Research Center, Air Research and Development Command.

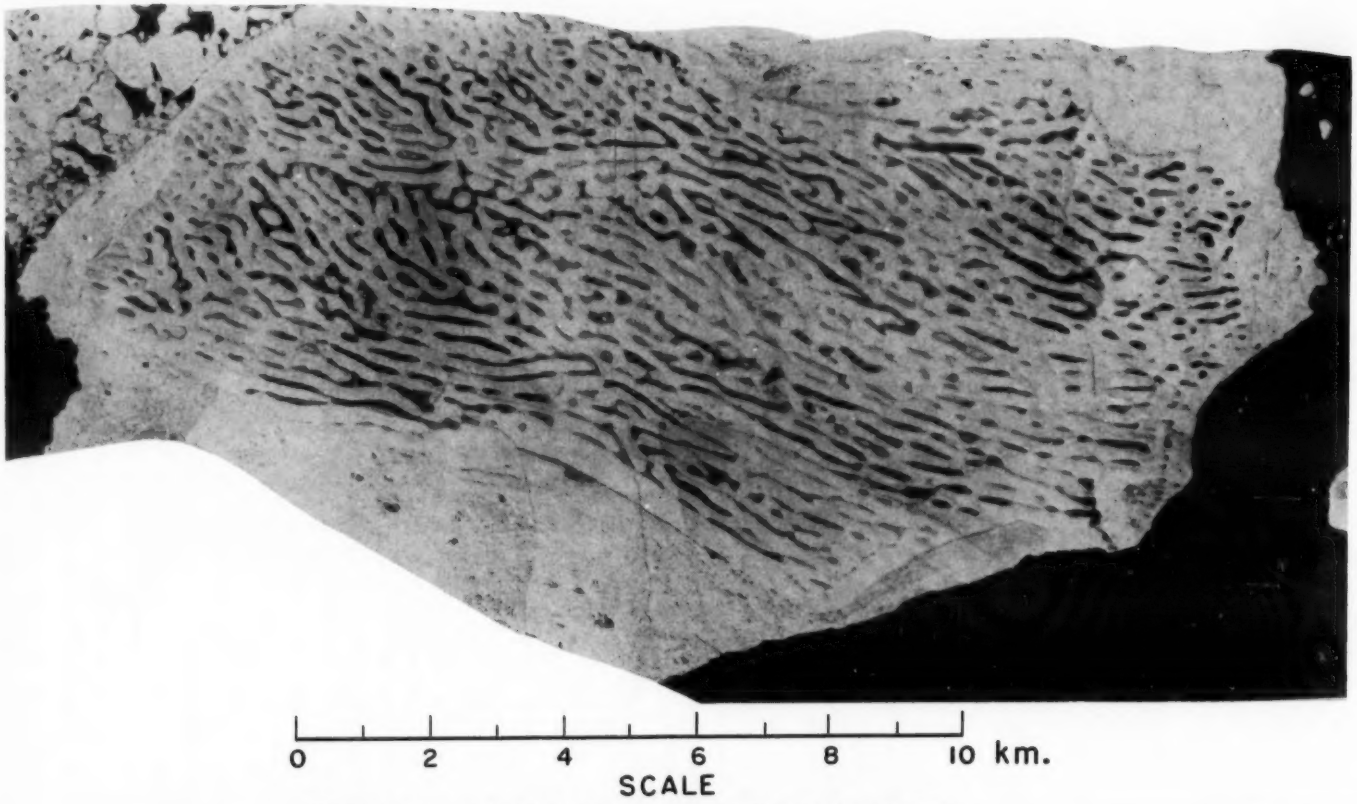


Fig. 2. Photo mosaic of ice island T-3. (By courtesy of U.S. Air Force).

The early history of the discovery of the arctic ice islands and the correlation with the ice shelf is given by Koenig *et al.* (1952). Résumés of the Air Force occupation and the early scientific programmes are given by Fletcher (1953) and Crary *et al.* (1952). Hattersley-Smith (1955) and Marshall (1955) have since described the general Ellesmere Ice Shelf programme.

The scientific operations in the ice island and ice shelf programmes fall into two major groups: the study of the arctic areas involved, their oceanography, geology, meteorology, etc., and the study of the ice island and ice shelf character. It is the intention to summarize in this article some general physical characteristics of the ice island and the ice shelf.

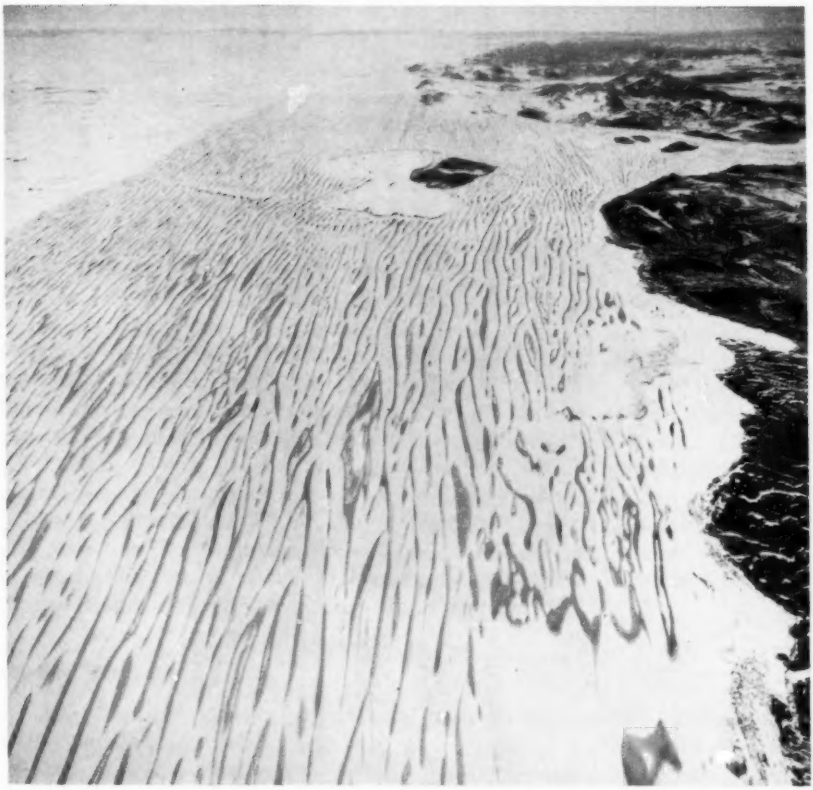


Photo: R.C.A.F.

Fig. 3. Aerial photo of Ellesmere ice shelf showing Ward Hunt Island.

SCALE

Fig. 2. Photo mosaic of ice island T-3. (By courtesy of U.S. Air Force).

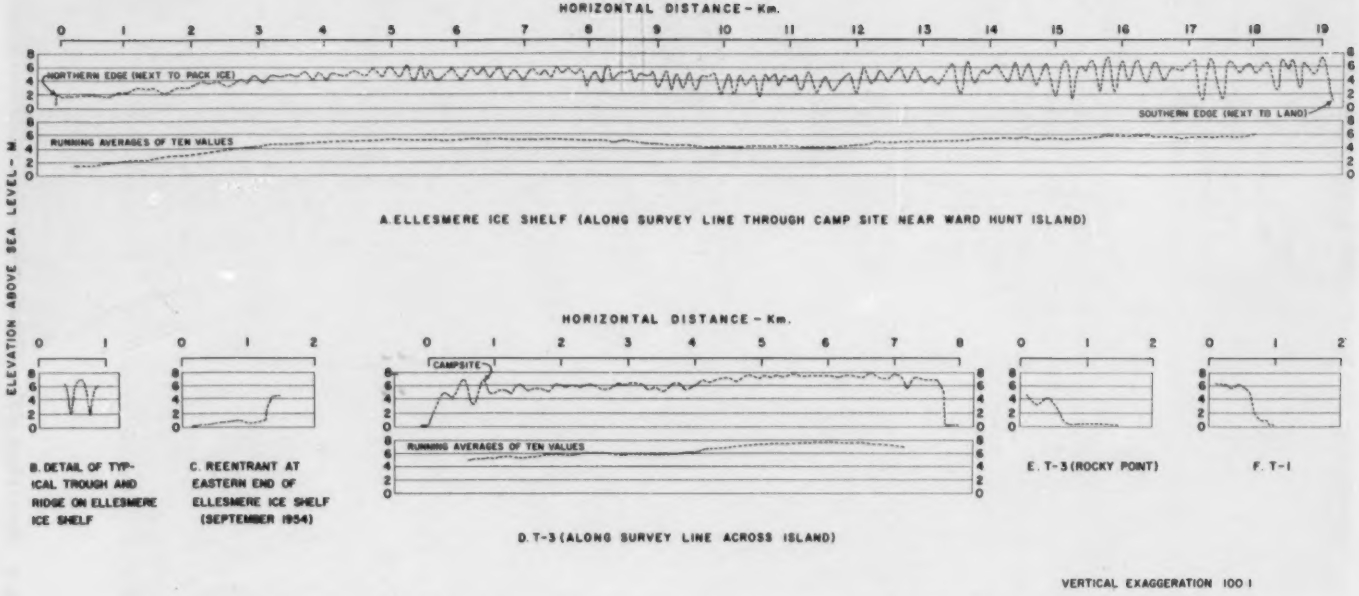

HORIZONTAL DISTANCE - Km.
NORTHERN EDGE (NEXT TO PACK ICE)
SOUTHERN EDGE (NEXT TO LAND)
RUNNING AVERAGES OF TEN VALUES
ELEVATION ABOVE SEA LEVEL - M
A. ELLESMERE ICE SHELF (ALONG SURVEY LINE THROUGH CAMP SITE NEAR WARD HUNT ISLAND)
HORIZONTAL DISTANCE - Km.
CAMPSITE
RUNNING AVERAGES OF TEN VALUES
B. DETAIL OF TYPICAL TROUGH AND RIDGE ON ELLESMERE ICE SHELF
C. REENTRANT AT EASTERN END OF ELLESMERE ICE SHELF (SEPTEMBER 1954)
D. T-3 (ALONG SURVEY LINE ACROSS ISLAND)
E. T-3 (ROCKY POINT)
F. T-1
VERTICAL EXAGGERATION 100 I

Fig. 4. Surface elevations of the northern Ellesmere ice shelf and of the ice islands T-1 and T-3

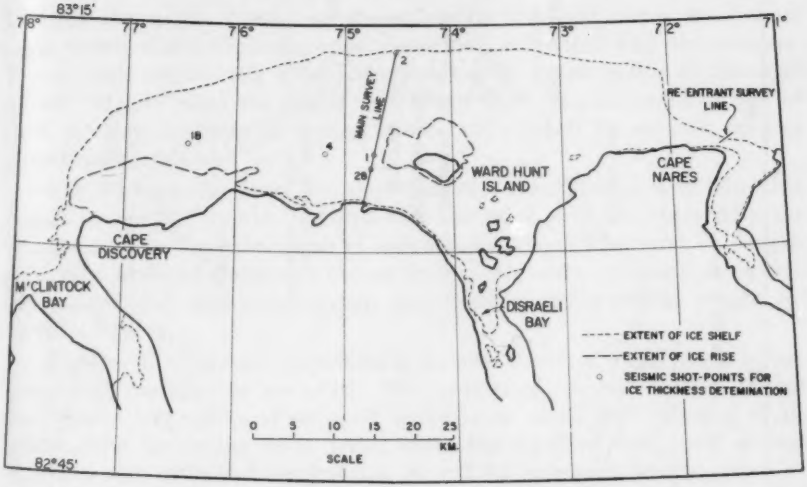


Fig. 5. Map of Ellesmere ice shelf showing the locations of survey lines and points where ice thickness was determined.

General configuration of ice island and ice shelf

Air photographs of the surface features of T-3, Fig. 2, and the Ellesmere Ice Shelf, Fig. 3, show island and shelf with similar ridge and trough features which are very evident in summer, when the troughs are filled with melt-water. Some of these features, both of the island and the present shelf, are seen to be regular; others, especially in the protected bays of the shelf, are quite irregular. The collective areal extent of the many known floating ice islands is about 2,500 square kilometres, somewhat more than the extent of the present ice shelf along northern Ellesmere. Most of the floating islands appear to have found the end or near-end of their journey in the Canadian Arctic Archipelago, while a few are still adrift in the perennial pack ice of the Arctic Ocean. It is believed that the process of ice shelf building is continuous, as many ice features are noted in aerial photographs, which appear less massive but have surface characteristics similar to the Ward Hunt Ice Shelf.

Surface elevations

Fig. 4 shows the results of transit level surveys across various parts of the ice shelf and island and Figs. 5 and 6 show the locations of these surveys. Values were taken to the hard ice surface under the snow cover.

The main survey across the Ellesmere Ice Shelf (Fig. 4A) shows the well developed ridge and trough systems on the inshore portion. The exact over-all extent of the area of little relief is not known though it is present along the entire extent of the west, north, and northeast edges.

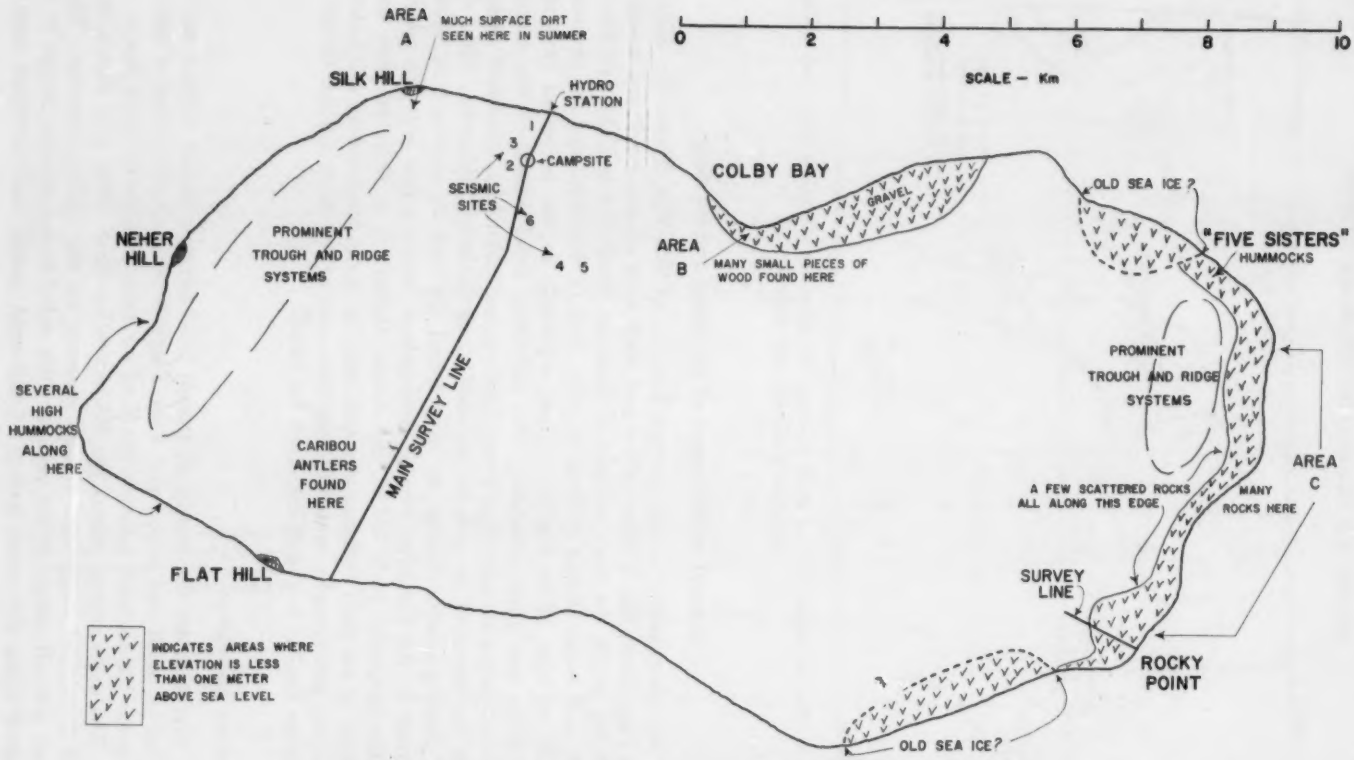


Fig. 6. Map of T-3 showing location of survey line and other points of interest.

There is also a low "moat" at approximately sea-level along some shore areas where warm drainage water from land collects during the summer. The ice shelf is obviously afloat throughout, with the exception of the small buried "islands" near the mouth of Disraeli Bay. The junctions of the ice shelf with the ice rises or buried islands are marked by obvious tidal or strand crack systems.

Fig. 4B shows in detail the elevations of a typical ridge with adjoining trough system. It should be noted that the ridge is much wider than the troughs, and in the calculation of average elevations this must be considered. The average elevations shown in Fig. 4A were obtained using five adjoining trough and ridge system elevations, allowing double weight for all ridge figures.

Figure 4C shows the elevations of an area north of Cape Nares immediately off the edge of the shelf. This area is of interest because it was here that a large piece of ice shelf broke loose about 1946 (Koenig *et al.*, 1952). The elevations were taken after the summer melt, and do not conform with estimated thicknesses, as will be discussed later.

Figures 4D and 4E are surveys made of the T-3 surface. The ridge and trough systems are not apparent on the main survey although the aerial picture shows the melt lakes very plainly. Figure 6 shows the approximate regions on T-3 where the ridge and trough systems are prominent, also the areas in which the elevations were probably less than 1 metre above sea-level. Elevations of one of these low lying areas are shown in Fig. 4E.

On T-1 the brief reconnaissance revealed no apparent ridge and trough systems. From aerial examination the surface of this large island is believed to have very little relief (Koenig *et al.*, 1952).

Fig. 6. Map of T-3 showing location of survey line and other points of interest.

Thickness and average density

Crary (1954) has described various seismic methods of measuring island thickness, three of which are dependent on plate waves. These three methods average the ice thickness over a distance of a kilometre or more. The fourth involves travel-time comparisons of seismic waves reflected from the ocean bottom in an area of uniform depths. In 1955 an additional thickness determination was made on T-3, using the horizontally polarized shear wave, (SH), reflected from the ice-water interface.

Attempts to obtain ice thickness on the Ellesmere Ice Shelf in 1954 were limited to six sites on the thick ice and one on the newer ice north of Cape Nares. On the ice shelf itself, at five sites, the SH shear wave was produced by a blow on a piton driven horizontally in the side of a pit. Detectors were located in a horizontal orientation and parallel to the force of the blow. Figure 7 shows a sample record in which four reflections of this SH wave to and from the top and bottom of the ice shelf are noted. The SH shear wave is not propagated in a fluid nor modified at the lower boundary into other wave types (Press and Ewing, 1951), and hence its total energy is reflected back from the bottom of the ice.

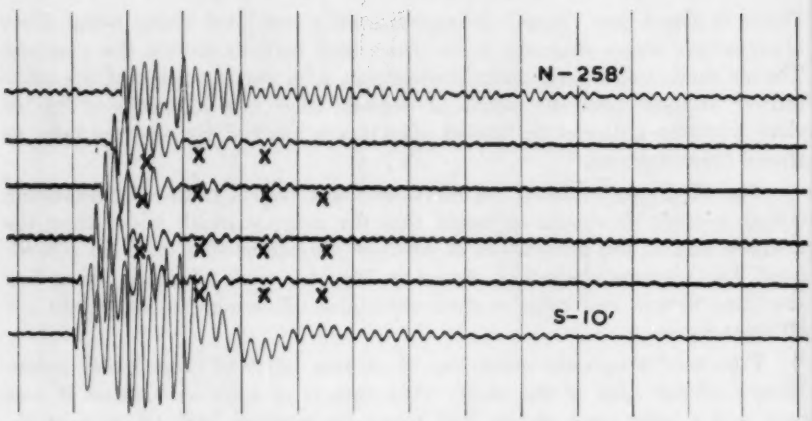


Fig. 7. Sample seismic record showing SH wave with four reflections.

In the calculations of thickness by this method, a slight modification of the reflection formulas was used. Repeated reflections between the bottom and top of the ice layer are described by the formula:

$$V^2\ T^2 = 4n^2\ H^2 + X^2,$$

where V is the shear wave velocity in ice, T the travel time, n the number of reflections from the bottom, H the thickness, and X the distance between the source of energy and the detector.

Although the reflection energies are quite evident, the initial impulse cannot be accurately timed except in a few cases. However, some identical phase of the several reflections can always be identified, and if we let the time difference between this part of the reflection and the actual beginning be e, the formula then becomes:

$$V^2(T - e)^2 = 4n^2\ H^2 + X^2,$$

where T now represents time to the identified part of the reflection. Comparing any two arrivals, with m and n number of reflections from the bottom occurring at the same distance, X, we get, by eliminating e:

$$H^2 = \frac{V^2\ TmTn(Tn - Tm) + X^2(Tn - Tm)}{4(n^2Tm - m^2Tn)}$$

In addition to the SH reflections the multiple reflected SV shear wave which had proved successful in the ice island thickness studies was used at one site on the Ellesmere Ice Shelf.

In order to obtain thickness values by these methods, it is necessary to know the velocity of shear waves in shelf ice. Figure 8 shows the time-distance refraction curve obtained from mechanical impulses, from which a velocity of 1,880 metres per second was obtained. Table 1 summarizes the various thickness determinations for the locations shown on the maps in Figs. 5 and 6.

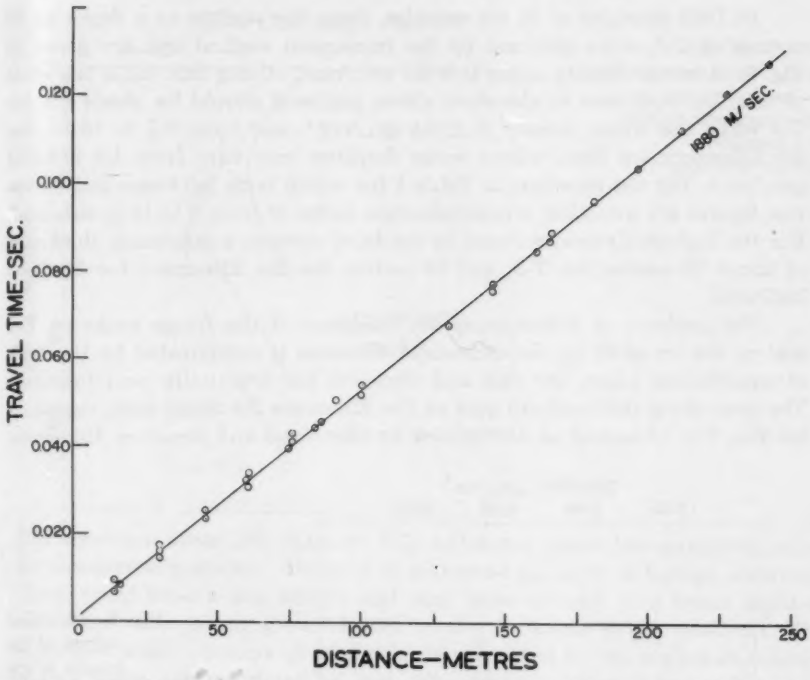


Fig. 8. Time-distance refraction curve giving velocity of shear waves in shelf ice.

Table 1. Values of ice thickness as determined by seismic means.

Location	*Seismic method used*	*Elevation (metres)*	*Thickness (metres)*
Ellesmere Ice Shelf			
Shot point No. 1	First SH reflection	4.5	45
" " No. 1	SH reflections	4.5	43
" " No. 3	" "	>4.6	51
" " No. 4	" "	>4.6	51
" " No. 18	" "	>4.6	53
" " No. 28	" "	6.5	52
" " No. 28	SV wave	6.5	54
Ice island T-3			
Site 1	Ocean bottom reflections	4.6	40
" 2	" " "	5.1	47
" 3	SH reflection	—	46
" 4	Flexural wave	—	50
" 5	Air-coupled wave	—	51
" 6	SV multiple wave	—	53

In 1952 densities of 29 ice samples, from the surface to a depth of 16 metres on T-3, were obtained by the immersion method and are given in Fig. 9. Average density value is 0.905 gm./cm.^{3}. Using this value the ratio of total ice thickness to elevation above sea-level should be about 9:1 for T-3 when the water density is 1.024 gm./cm.^{3}, and from 9:1 to 10.5:1 for the Ellesmere Ice Shelf where water densities may vary from 1.0 to 1.024 gm./cm.^{3}. For the locations in Table 1 for which both thickness and elevation figures are available, a multiplication factor of from 8 to 10 is obtained. For the highest elevations found in the level surveys a maximum thickness of about 68 metres for T-3, and 60 metres for the Ellesmere Ice Shelf is indicated.

The problem of determining the thickness of the fringe areas on T-3 and on the ice shelf by elevations and densities is complicated by the lack of equilibrium where the thin and thick ice are essentially held together. The area along the northern part of the Ellesmere Ice Shelf must certainly be thin, 9 to 12 metres as determined by elevations and densities. However,

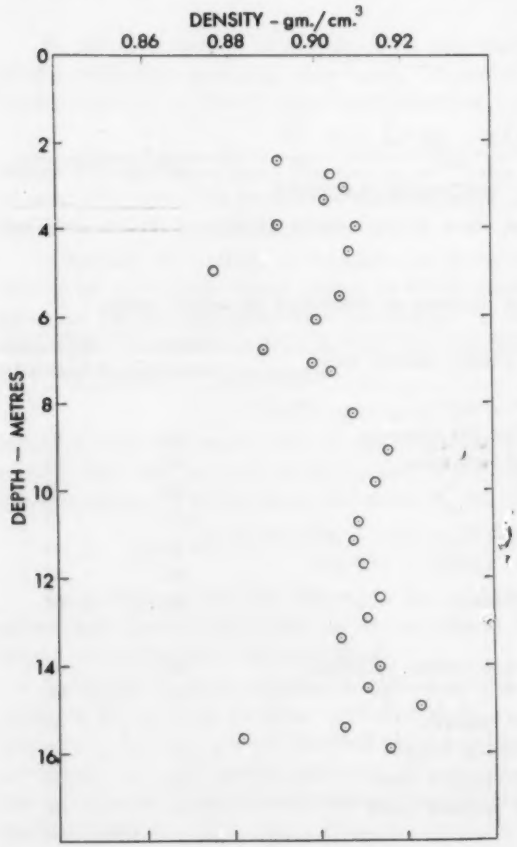


Fig. 9. Observed values of ice density in ice island T-3.

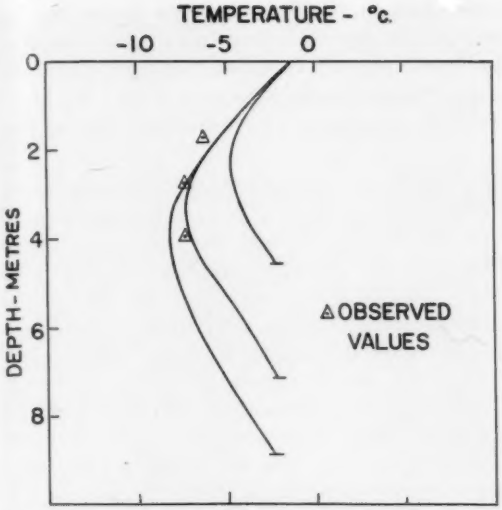


Fig. 10. Fit of theoretical temperature curves for ice of varying thicknesses to observed ice temperature values.

low elevation areas like those on T-3, which are much less extensive, may be considerably thicker. Because of abnormal amounts of foreign material, these areas have a low albedo and may have ablated to a lower surface elevation without any great change in the thickness below sea-level. On a smaller scale the same phenomenon is presented by the melt-water lakes where these are restricted in area. The surfaces of melt-water lakes of many of the troughs inside Disraeli Bay are at sea-level.

An unsuccessful attempt was made in September 1954 to determine the ice thickness by using seismic flexural waves in the area where a portion of shelf had broken off in 1946 north of Cape Nares. It is believed that the failure was due to the variations in ice thickness caused by the melt-water streams. Holes were drilled to a depth of 6 metres, where salt water infiltration prevented further deepening. An attempt was made to determine the thickness from observed below-surface ice temperatures, using an assumed value for surface temperature variations. Fig. 10 shows three theoretical temperature curves computed for depths of 4.5, 7.0, and 8.9 metres. A depth of 7 metres shows the best fit to the observed temperature values. With consideration of known values of the average temperature range, thermal coefficients of the ice, and the validity of the assumption of negligible surface change, an ice thickness of 7 metres is approximately the value for the growth during 7 years.

Surface features and surface deposits

Surface dirt layer

One of the most obvious surface features of the ice island and ice shelf is the thin layer of dirt near the surface. This dirt was not distributed uniformly but collected in small holes generally from a few millimetres

to about 2 centimetres in width and up to 10 centimetres below the ice surface. The small collections of dirt, by absorbing more radiation than the ice, have melted out the small deposit holes and caused continued melting ahead of the ice surface. Some individual grains of dirt remained on the surface to give it a darkish appearance after summer melting had removed the snow cover.

The amount of this "surface" dirt layer was determined in two instances in shallow pits approximately 1 square metre in area near the campsite at T-3. Values obtained were 114 and 122 gm./m.^{2}. At the campsite location on the ice shelf in northern Ellesmere a similar determination gave a value of 45 gm./m.^{2}. If evenly spaced, the T-3 dirt layer would be only of the order of 1/20 of a millimetre thick. It is difficult to estimate the fraction of this dirt that may be lost by drainage during summer melting but ordinarily the dirt is protected by its location in the small holes. An exception to this was noted during a very heavy rainstorm on the northern Ellesmere Ice Shelf in mid-August 1954, when the dirt in many places was literally flushed from its protecting beds. The dirt layer did not seem to vary in appearance anywhere on the high areas of the ice island, but on the ice shelf it was less noticeable on the seaward edge than it was over the shoreward areas of the shelf. At T-1 there was the same type of dirt layer in the area visited, though perhaps not as heavy as on T-3.

Examination of the mineral content of the dirt on T-3 has been made by Stoiber, Lyons, Elberty, and McCrehan of Dartmouth College (1956). They report the following estimated percentages of minerals in the surface dirt: 60% lithic fragments (fine-grained altered volcanics), 5% quartz, 10% plagioclase, 5% muscovite, 10% biotite, 5% K-feldspar, with small amounts of hornblende, chlorite, sphene, calcite, garnet, etc. They found "no essential mineralogical difference between the surface dusts of T-3, T-1, and the dusts from within the ice of T-3. It is also evident that the dusts are mineralogically identical with the rocks collected at the surface of T-3." They appear to have been derived from a land mass underlain partly by volcanic rocks and partly by metamorphic and plutonic types. The low percentage of heavy minerals in the dust makes it probable that it was wind-deposited. It was noted that the low-lying hills adjacent to the coast near the Ellesmere Ice Shelf were snow free from mid-June until about mid-August. Prevailing coastwise winds in that area would account for the decrease in the amount of the deposited layer toward the seaward edge of the shelf.

This surface dust also contains some organic material. Dr. Barghoorn of Harvard University (personal communication) has kindly given the following description for one of the samples. "Organic residues in the sample consist of a considerable range of microfossils. These include threads of fungus hyphae, algal (?) spores, fungus spores, and fairly numerous fragments of woody tissues of higher plants, commonly infested with fungus hyphae ; resinous bodies and a very few pollen grains." However, he found that "other samples collected in the summer of 1955 exhibit a very low proportion of organic matter and a high proportion of acid

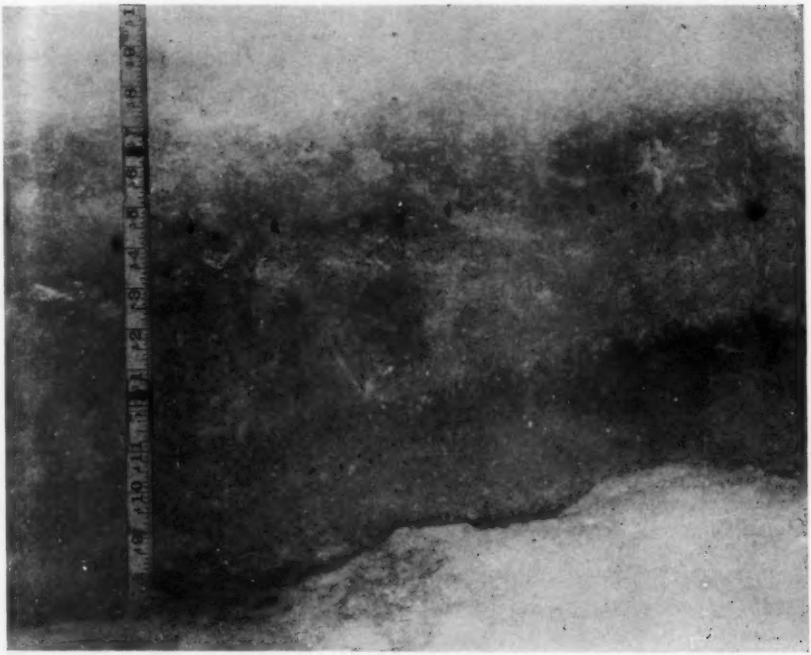


Fig. 11. Section of T-3 ice showing heavy top dirt layer and one lower layer.

insoluble minerals. Standard procedures for extraction of pollen and spores yielded negative results and the only organic constituents found were amorphous, finely divided fragments of particulate organic matter. A few discrete cells of probable algal origin were observed."

Very soon after the occupation of T-3, as excavations were made into the ice, it was noted that in addition to the top surface dirt layer there were other, much lighter layers in the interior of the ice. Fig. 11 shows the dirt in the top layer, and one of the heavier of the lower dirt layers. Some of these layers were extremely light, so that the grains were actually separated. It was quite apparent that the top layer must represent a long period of collection either by continual melting into the ice surface or by the surface remaining virtually unchanged for many years. In no locations were there thin dirt layers buried in the accumulated snow or ice above this heavy layer. Therefore it must have been many years since snow had accumulated in the area of its origin. On the Ellesmere shelf during the summer of 1954, which was an exceptionally long summer, the dust deposited was not noticeable. It is very unlikely also that more than one light layer could be deposited annually since the low-lying hills are bare for only about two months. Summer snowfall heavy enough to cover the ice would also cover these hills and the melting in one area would be closely paralleled by melting in the other.

Plant and animal specimens

During the period in which T-3 has been occupied, about one-half of the island surface has been examined. This was possible only during the late summer period after the snow cover had melted. Although the entire surface of T-3 was characterized by a honeycombed appearance with the small aggregations of dirt of the top layer at the bottom of the holes, there were several areas of particular interest as indicated in Fig. 6. In the area "A", about 2 to 3 square kilometres in size, lying at or near sea-level, very large amounts of surface dirt were present; they were several times larger than those of the normal surface layer, and in many areas the dirt completely covered the ice as a thin mantle. In this heavy dirt layer were found a few fragments of plants, some lemming remains, and caribou excrement. Following the island edge clockwise, two fish skeletons were found about 2 metres above sea-level. These had broken apart and were dispersed in holes in the honeycombed ice. It was seen that in most of the shore areas along this part of the island where elevations change to sea-level gradually rather than dropping off abruptly, the holes containing the dirt layers were much larger, in many cases up to 30 or 40 centimetres in diameter.

In area "B" large amounts of plant material were located. Dr. Polunin (1955) made a thorough examination of this material and the following information is obtained from his report. "Plant materials collected during 1952, 1953, and 1955 at or near the surface of the ice island T-3 belong mainly to arctic willow, *Salix arctica,* (531 items, mostly pieces of stems or roots up to 15 centimetres long), purple saxifrage, *Saxifraga oppositifolia* (164 items, including tussocks up to 12 centimetres in diameter), low spear grass, *Poa abbreviata* (20 partial specimens, including a tussock 9 centimetres in diameter), and tufted saxifrage, *Saxifraga caespitosa* (7 small rosettes, etc.). In addition, there were determined various leaves, roots, and faeces; also four species of lichens and fairly numerous mosses and liverworts of which some were represented by sizeable tussocks. All were dead except a circular mat 24 centimetres in diameter of the moss *Hygrohypnum polare,* which proved to be still alive. Gathered during the protracted drift of the ice island in the Polar Basin, and in many cases in the vicinity of the North Pole, these land-grown botanical items with a few understandable exceptions strongly suggest very high-arctic origin, and it is supposed that they were washed (or possibly in some cases blown) on to the surface of T-3 when it formed part of the shelf-ice attached most likely to the north coast of Ellesmere Island. So far as this last region has been investigated, available data would seem to support this suggestion." Most of the area "B" was near sea-level and also had an unusually large amount of surface dirt. Near the far end of area "B" some gravel deposits were located. A thorough examination of this area was planned for 1955 but for the first time during the occupation the snow never completely melted and surface studies were not possible.

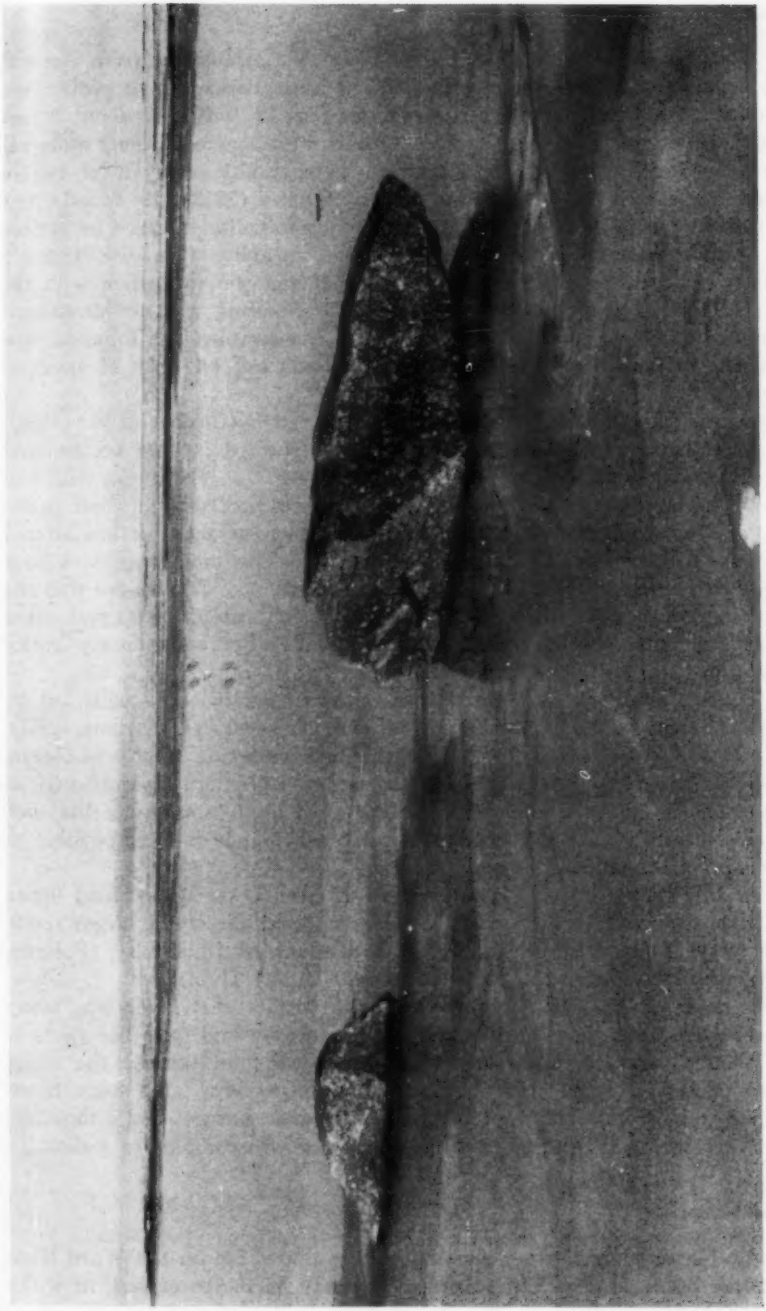


Fig. 12. Large rock near area C on T-3 (length approximately 3 feet).

Geological specimens

Starting with an isolated specimen about 2 kilometres from Area C and continuing around that end of T-3, erratic rocks and gravels were common. The isolated specimen shown in Fig. 12, which is about 3 feet long, was perhaps one of the largest. Others were located in long morainal features over 1 kilometre in length. A petrographic analysis of twelve rock samples collected in 1952 was done by Shorey (1953). He found "four gneissoid rocks of granitic composition, two amphibolite gneisses, an actinolite granite gneiss, a granite micropegmatite, an aplite, two felsites, and a diabase. The gneissoid rocks appear to be of metasomatic origin with the mafics delineating the relict bedding planes. Banding was not prominent in any of the specimens and in only two of the samples was foliation outstanding. The minerals found in the gneisses are all only of medium grade."

A total of 153 rocks from T-3 was examined by Stoiber *et al.* (1956). Of these, "a representative group of 38 was studied in thin section and the remainder were examined in hand specimens. . . . The rocks with few exceptions may be divided into several categories according to their petrography. The general subdivisions are schists, gneisses, more or less altered extrusives, intrusives, (and) graywackes. . . . The most abundant rock type represented in the suite is altered volcanics. . . . It is noted that the following rock types are not present: slates, chlorite schists, and other low-grade metamorphic rocks; limestones and other sedimentary rocks, except for two graywacke specimens."

From a comparison of these rock specimens with those collected by Christie in his study of northern Ellesmere shore areas (Christie, 1957), Stoiber *et al.* (1956) conclude that "of the land bordering the Arctic Ocean, areas on the north and northwest coast of Ellesmere Island such as Yelverton Bay and Cape Bourne are the most likely sources of the rock material on T-3." On the basis of available evidence it is impossible to fix the precise location of origin.

In 1955 in the general rocky terrain of Area C on T-3, a dead lichen specimen was collected by Norman Goldstein from one of the larger rocks. This was examined by Dr. I. M. Lamb of Harvard University (Polunin, 1955), who identified it as *Caloplaca elegans* (Link) Th. Fr.

In most of Area C, as in Area A, the surface dirt layer was many times heavier than had been noted at the campsite and on other parts of the island. There was also some animal material found among the rocks. This included a small fish about 10 centimetres long and some bones thought to be from a polar bear. Area C, however, has not had a thorough searching during the period in which the snow was completely gone.

Strand cracks

At the junction between floating and grounded ice on the Ward Hunt Ice Shelf occur obvious strand cracks, usually ½ metre or less in width

A. STRAND CRACK RECORDS
CRACK #1

Fig. 13. Recorded amplitudes of strand crack opening and closing; water-level observations; and bubble level readings (Ellesmere 1954).

and 2 to 8 metres deep. These cracks were not continuous along the outer fringe of the grounded zone but often occurred in "en echelon" formations. These were found around all buried islands, ice rises, and shore areas bordering the ice shelf, and were not observed elsewhere. These cracks, as will be shown, are the surface expression of strains produced by ocean tides.

In June and July 1954 close observations of actual tide levels were taken by R. L. Christie of the Geological Survey of Canada to whom the author is indebted for this information. His figures were obtained in open water in M'Clintock Bay and near Ward Hunt Island and are shown in Fig. 13C. The maximum tidal range appears to be about 30 centimetres and the diurnal tides show considerable inequality of range. The tide character appears generally similar to that at Resolute Bay and Isaachsen as given by Rae (Manning and Rae, 1950).

Observations were made of the opening and closing of the strand cracks in the western section of the Ward Hunt Island ice rise. This was done by freezing lever arms in the ice which operated a pencil over a spring-driven clock drum. In Fig. 13A are shown observations made during May, July, and again in late August and September. During the summer months when the shelf surface temperature is higher, the ice apparently can undergo considerable bending without fracture. The renewal of larger tide motion in the crack zone, shown in September, coincided with the appearance of a new small surface-crack entirely independent of the previous one which had filled with water and frozen. This new crack, though running closely to the older one, alternately was on one side and then the other and fortunately crossed the old crack underneath the recording mechanism. Comparisons of the opening and closing of the strand cracks with tidal observations during mid-July, when these observations are available in the same vicinity, show that the cracks tend to close at high tide as would be expected. The maximum amount of horizontal movement occurred during late May with a value of about 1 centimetre. Since the ice shelf thickness is about 40 metres in this vicinity, this would represent a surface tilt difference of about 100 seconds across the strand crack, if equal strains take place on the upper and lower ice surfaces.

To get further information of this ice tilt that is taking place near the grounded zone, bubble levels of 30 and 60 seconds resolution were frozen in the ice on either side of the stand crack. Figure 13B shows the results, which were obtained only for a few days of late August and early September, at a time when strand crack opening and closing was about a quarter to a third of that observed in May. Though an average range of 100 seconds was obtained, no particular difference was noted on opposite sides of the crack during the observational period. An appreciable tilt probably occurs as far back as 1000 metres from the grounded area, since no difference in tilt was observed over a distance of 100 metres.

Figure 14 shows the western half of the Ward Hunt ice rise. Along the western edge of the ice rise the fracture zone lies mainly on the seaward

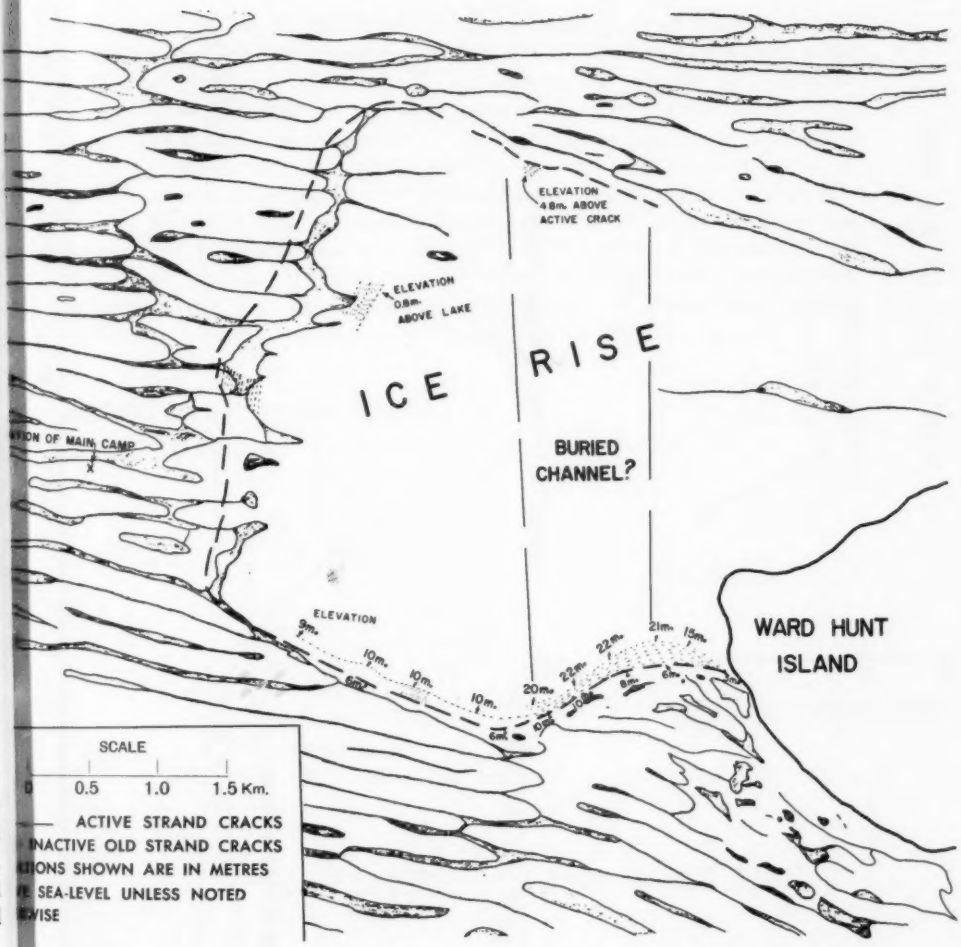
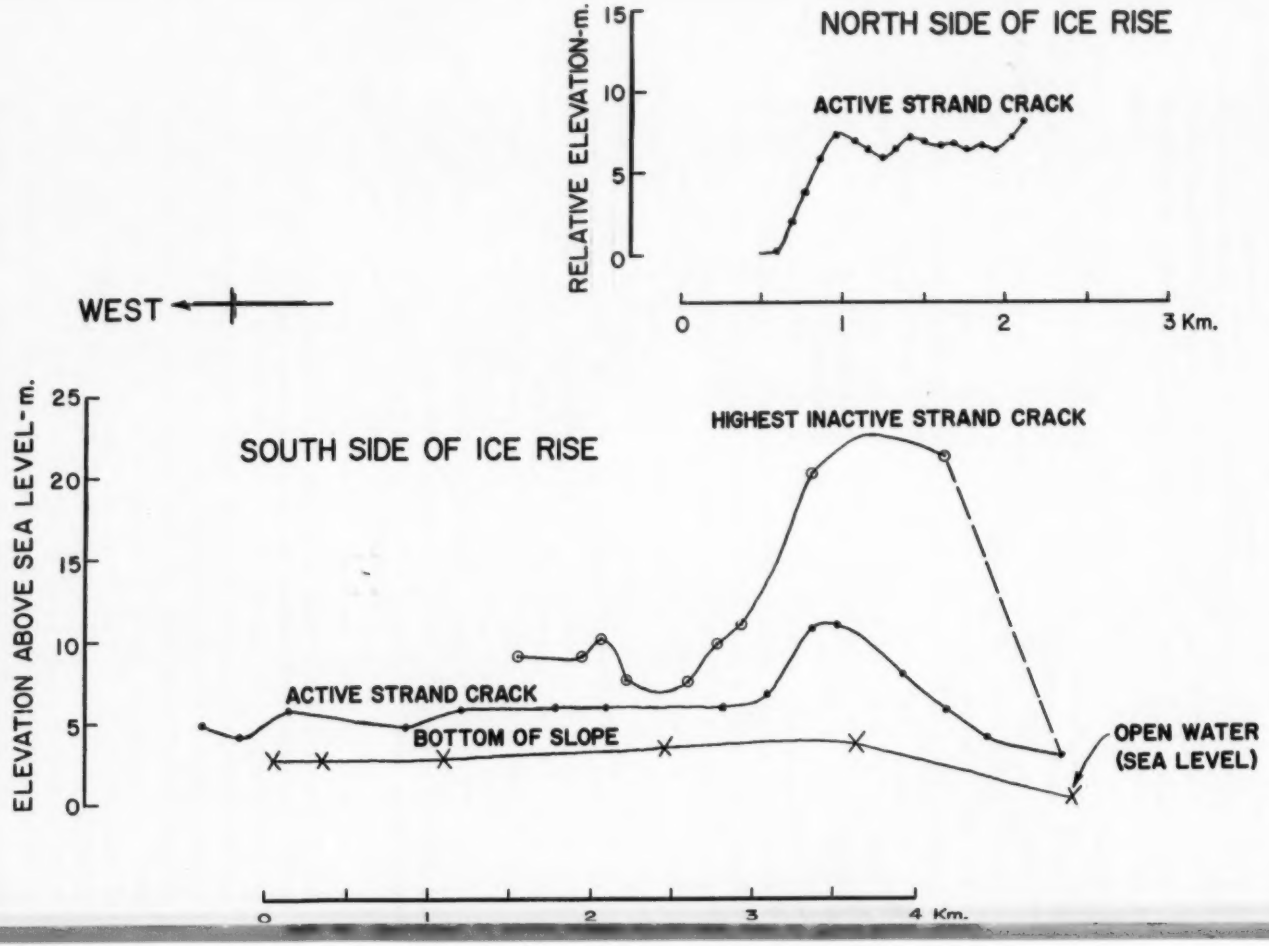


Fig. 14. Western portion of Ward Hunt ice rise and surrounding shelf ice. Location of the main active strand crack is shown, as well as many of the old inactive ones.

side of the lower elevation drainage lakes, while on the northern and southern edges it lies inshore and parallel to the lakes, in places rising up on the slopes of the ice rise. The elevation of the fracture zone above sea-level will necessarily be an indication of the thickness of the ice which is balanced on the bottom, hence the changes in elevation will be an indication of the sub-ice topography except where the strand cracks occur in narrow deep channels, lakes, or other minor surface drainage features.

Figure 15 shows the approximate elevations above sea-elevel of the strand crack south of the ice rise and elevations relative to lake level of


NORTH SIDE OF ICE RISE
RELATIVE ELEVATION-m.
15
10
5
0
ACTIVE STRAND CRACK
0
1
2
3 Km.
WEST
SOUTH SIDE OF ICE RISE
ELEVATION ABOVE SEA LEVEL-m.
25
20
15
10
5
0
HIGHEST INACTIVE STRAND CRACK
ACTIVE STRAND CRACK
BOTTOM OF SLOPE
OPEN WATER
(SEA LEVEL)
0
1
2
3
4 Km.

the crack to the north. Sea-level elevations were obtained from the north-south line through the campsite (see Fig. 5) and checked at the open water southwest of Ward Hunt Island. An interesting feature is a possible channel under the ice rise. This is suggested by the higher elevations of the strand cracks, as indicated. This channel, perhaps over 60 metres below sea-level, may be responsible for the pronounced saddle in the ice rise surface immediately west of Ward Hunt Island. At another smaller ice rise south of the ice shelf and immediately off shore just west of Disraeli Bay, the presence of a channel is also suggested by strand crack elevations. This is supported to some extent by bathymetric information, as seen in Fig. 16. This channel is immediately off shore from a small stream now diverted eastward to the edge of the ice rise.

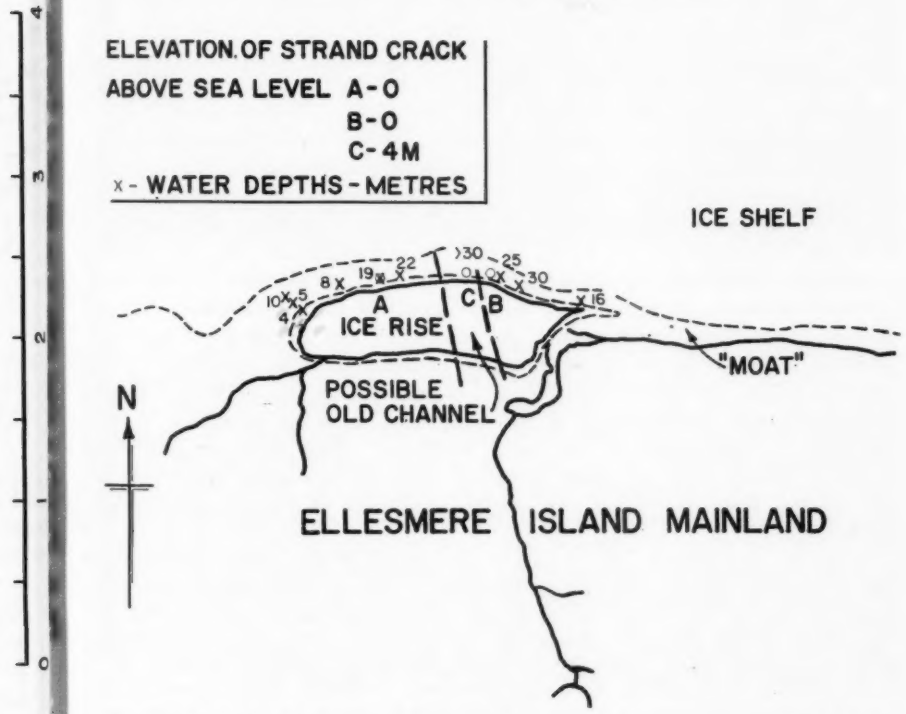


Fig. 16. Water depths and strand crack elevations at small ice rise west of Disraeli Bay.

Of more importance in a historical study of the ice shelf and ice island are the old scars left by strand cracks of previous periods. Fortunately, the 1954 summer was quite prolonged and allowed studies after the snow melt was complete. As noted above, the strand cracks are considered as expressions of strains in a zone that must increase in width with a smaller angle of dip of the ground surface under the ice. During the

summer seasons strand cracks fill with drainage water which freezes before lower surface temperatures start another fracturing cycle. However, through the years many factors influence the fracture-zone location: increase or decrease of the thickness of the floating ice, changes in sea-level, or rise and fall of land areas due to isostatic adjustment. Hence, the older fracture scars that are no longer in the tidal adjustment zone are evidences of past history. Fortunately, the ice frozen in these strand cracks differs sufficiently in character from the normal ice of the shelf formed from accumulated snow so that they are quite evident, as shown in Fig. 17. Figure 14 shows the general areas of old strand cracks. Figure 15 shows elevations of the highest inactive fracture on the southern edge of the ice rise. In some locations these older fractures were as high as 10 to 15 metres above the present zones and in all but a few cases they were on the ice that is grounded at present.

Considering the possible causes of variations in strand crack location, accumulation or ablation from the surface causing a change in the ice thickness would move the balance zone seaward or landward respectively. These cracks would always develop at the floating ice surface, so that the elevations of the cracks would vary only by about one-ninth of the total change in ice thickness. Also, general accumulation and ablation to a marked degree would bury or erase by thawing old strand crack evidence. A recent, relatively short thawing period could explain the very few areas of fracturing seaward of the present active cracks. The absence of extensive fractures seaward, however, may be the result of the simultaneous rising of land areas that would nullify the effects of the thinning of the ice. Another possible change, which must be considered, is the building up of ice from below. Build-up from below without surface change would move the active crack seaward and at the same time raise the elevation of the fractures by one-ninth of the total thickness. This may explain the presence of the lower elevation drainage lake systems along the western part of the Ward Hunt ice rise on the land side of the present strand crack system.

For an explanation of the higher elevations of the older systems on the north and south side of the ice rise, it seems most likely that an isostatic rise of the Ward Hunt Island group has taken place. This is confirmed by the presence of marine shells on the raised beach of Ward Hunt Island, for which carbon-14 studies give an age of 7200 years.

Subsurface observations

Shallow sections

In August 1952 ice cores averaging 1 metre long were obtained by drilling along several different sections in the vicinity of the campsite at T-3. These sections were located over the ridges only, since drilling was difficult through the water in the troughs. The purpose of obtaining these shallow sections was to learn more about the light dirt layers; their con-

Fig. 17. Systems of old strand cracks in the ice shelf near the Ellesmere mainland.

tinuity, consistency, and attitude with respect to the heavy dirt layer immediately at the surface. Figure 18 shows the locations and Figure 19 shows the profiles. No studies were made of the ice character of these cores. It is noted that although there is general continuity to the dirt layers it is by no means evident in all cases. In many of the profiles, holes had to be bored 1 metre apart to establish correlation. The deformation of these dirt layers can be attributed only to a nearly stagnant ice body existing over a long period of time. The second noticeable feature is that there is a very definite discontinuity in all profiles between the top heavy dirt layer and the lower ones, which are themselves generally conformable. Figure 18 shows possible surface changes between the present time and the time in which the lighter layers underneath were deposited. This migration is not uniform, however, either in direction or distance. In general the main outlines of the ridges are not significantly different from those of the present.

As noted above, the amount of dirt in the top layer was greater by a factor of about ten, than that of any of the lower dirt layers noticed near the surface. The lighter layers also varied widely. Their weight per square metre ranged from about 12 grams for the lower one illustrated in Fig. 11, to less than 1 gram per square metre, which could not be easily measured, for many others. In many of the lower layers the same features as found in the top layer were noted on a smaller scale. Very fine dust,

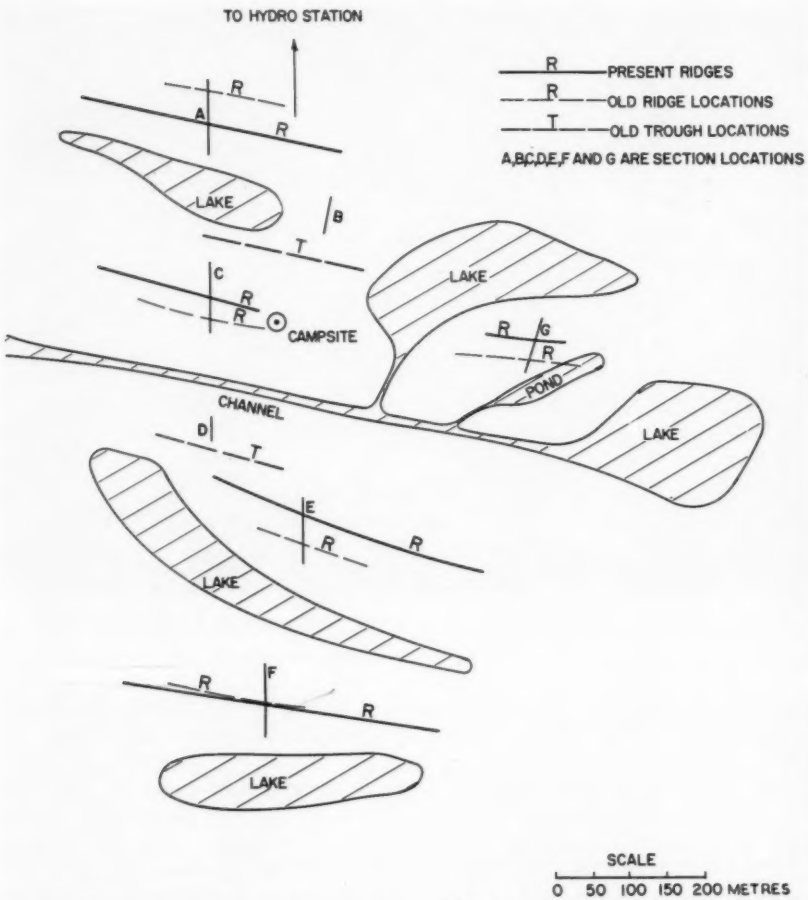


Fig. 18. T-3 campsite area showing locations of shallow profiles; inferred change in trough and ridge locations is indicated.

visible usually as discoloration, was followed within less than 1 centimetre by the masses collected in symmetrical holes, which had filled with melt water and frozen. Some grading was also noticeable, the larger masses being lower in the ice. Such layers must represent the result of many years' accumulation. The sections shown in Fig. 19 represent only the obvious dirt layers, those that were noticeable when the solid core was laid on the snow and washed with water. Later more careful examinations of cores taken in the vicinity of Section A revealed a few other, very much lighter layers that had been overlooked.

On the right-hand side of Sections C, E, and G, in Fig. 19, it was noted that the dirt particles were not in regular layers but scattered through

THIS CHART WILL BE THE FIRST CHART
APPEARING AT THE END OF THIS FILM

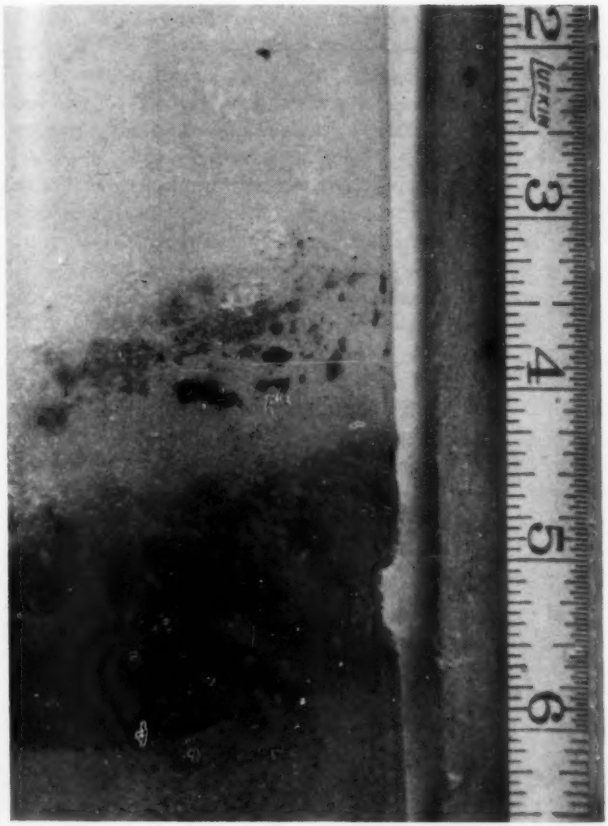


Fig. 20. Section of T-3 ice core showing heavy bottom dirt layer.

the ice or in many cases arranged in short, vertical lines. It is believed that this is the result of deposits on lake ice where crystals are usually elongated vertically. When thawing progresses on such ice, crystal boundaries melt first, allowing the dirt particles to penetrate between the crystals. Such conditions were observed at the surface of the ice shelf in northern Ellesmere in 1954 and were interpreted as indicating locations of former lakes or ponds.

It is noted that some migration of trough and ridge systems appears to have taken place since the lower dirt layers were deposited.

Deep drilling

Dirt layers. A total of nine deep holes for study of the dirt layers and ice character was drilled on T-3 on the campsite ridge and across the intervening trough to the adjacent ridge between Sections A and B, Fig. 18. The first deep hole was drilled in August 1952 to 15.8 metres. Throughout this length of core continuous dirt bands were encountered, totalling nearly

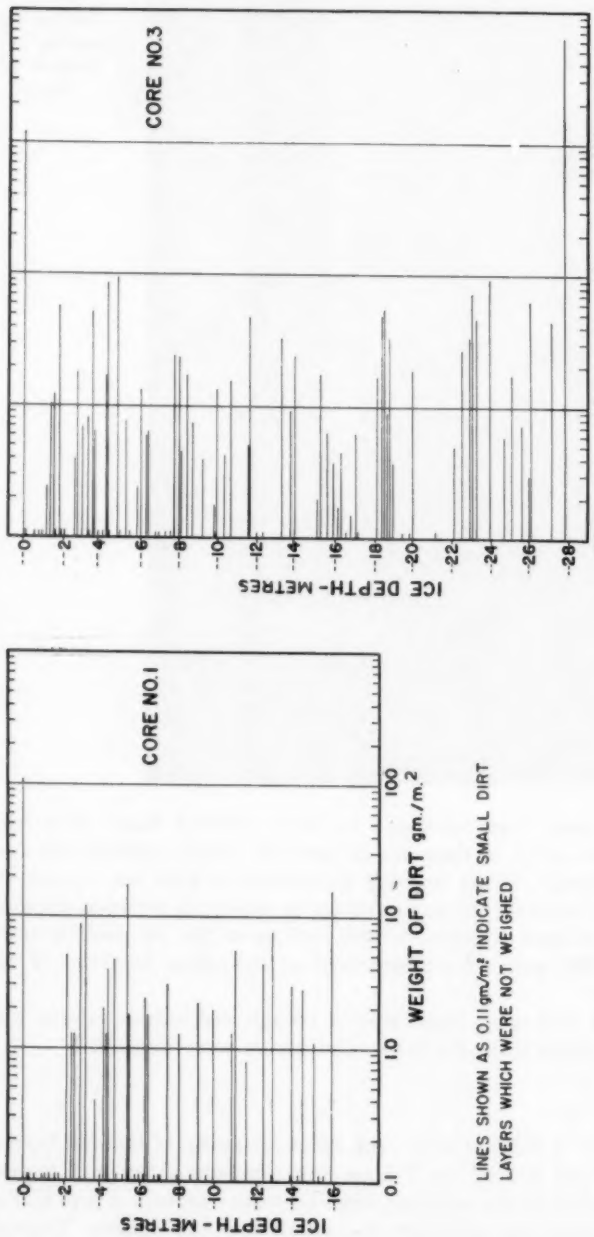

CORE NO.3
ICE DEPTH - METRES
WEIGHT OF DIRT gm./m.²
CORE NO.1
ICE DEPTH - METRES
WEIGHT OF DIRT gm./m.²
LINES SHOWN AS 0.1 gm/m² INDICATE SMALL DIRT LAYERS WHICH WERE NOT WEIGHED

Fig. 21. Depths and weights of dirt layers as determined by two T-3 cores.

50. In all cases, however, none were nearly as heavy as the layer at the surface. In 1953 two other deep holes were drilled, Hole 2 on the same campsite ridge as Hole 1, and Hole 3 on the opposite ridge at the site of Section A. Hole 2 was used for ice petrographic study (Marshall, 1955) and was drilled to 32.5 metres. Dirt layers were noted down to 25.4 metres where a very heavy layer was encountered. Figure 20 illustrates this heavy layer. Below this in the remaining 7.1 metres of the hole, the ice was completely free of any dirt particles. In August 1953 Hole 3 was drilled to 28 metres in order to obtain the weights of the dirt layers. The heavier dirt layer was encountered at 27.8 metres, and 0.2 metre was cored into the dirt-free ice below. The weights of the heaviest of the dirt layers, extrapolated from the 8-centimetre-diameter core to 1 square metre, are given in Fig. 21 for Holes 1 and 3. As in the shallow sections, there was a lack of correlation between the cores as regards the heavier of the intermediate layers, which should themselves represent ablation periods of many tens or perhaps hundreds of years. The lower part of Hole 3 below approximately 20 metres consisted mainly of what has been defined above as lake ice according to the manner in which the dirt had collected, usually in vertical lines or individual masses instead of lying in layers such as had been found above.

Soon after Hole 2, which showed the extremely heavy bottom dirt layer followed by dirt-free ice, had been drilled, Marshall located outcrops where this layer had combined with the surface layer near the shore at Area A and later at Area C, Fig. 6. In these areas, this heavy layer was identical both in appearance and in lack of underlying dirt layers to that found in the deep holes. Thus, the heavy surface layer in those areas represented the full amount found in the total column at the campsite. Near these outcrops it was possible to obtain an average weight over a fairly large area of several square metres, and a value of 614 grams per square metre was obtained. Assuming uniform annual deposition this would represent about five times the melt period of the top surface dirt layers; and also from an assumption that the lightest dirt layers represent an annual accumulation, the conclusion is drawn that the outcrop deposition must have taken at least a span of time of the order of several thousand years.

In 1955, before the melt season had begun, a series of six deep holes was drilled across the valley between the ridges where the first three holes had been located. Figure 22 shows the results of this section and of the previous drilling. The heavy dirt layer located at depths 25 and 28 metres under the ridges was unexpectedly encountered at shallow depths in the 1955 drilling. Thus a complete migration of trough and ridge appeared to have taken place, though it is not evident if the distances between ridges were different though they cannot have been smaller, and appear not to have been much greater.

A third fairly heavy dirt layer was located immediately above the heavy bottom one in several holes along the top of the buried ridge. The

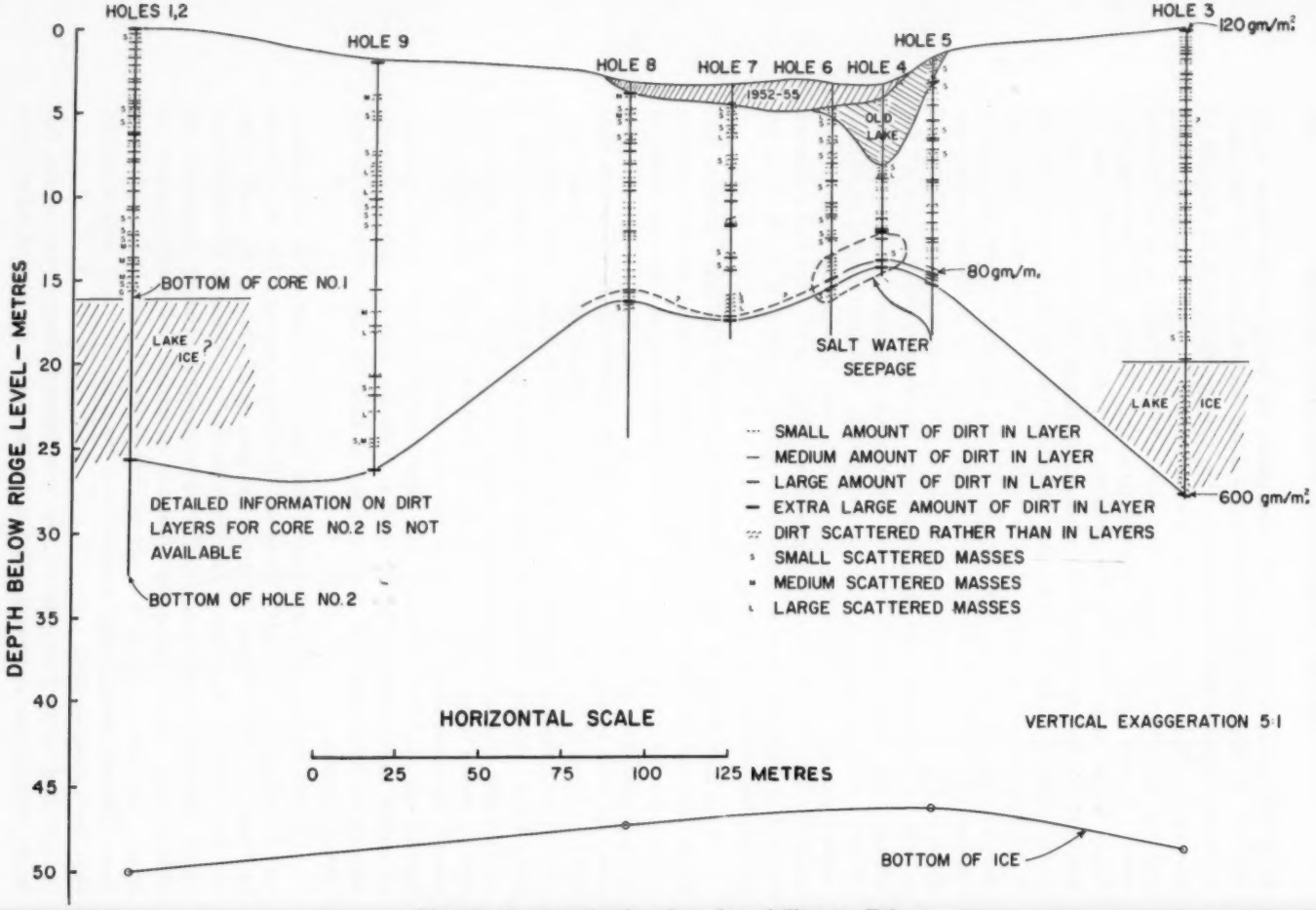


Fig. 22. Cross section based on deep drilling on T-3.

weights in two cases gave about 80 grams per square metre for this layer, not too different from that of the top one, and hence must represent another long period of ablation. The accumulative weights from the bottom dust layer to the top layer are shown in Fig. 23 for the cores of the present trough and ridge. The horizontal axis is in effect a time scale, providing dust was uniformly deposited. Previous periods of ice accumulation such as have formed the top part of T-3 probably existed prior to the ablation periods in which the dirt of the two lower layers was collected.

A comparison of the lower dirt layers including the heavy bottom layer was made by Stoiber *et al.* (1956) and showed that no differences whatever existed as regards the mineralogical content, grain size, or percentage of heavy minerals between any of the dirt layers encountered.

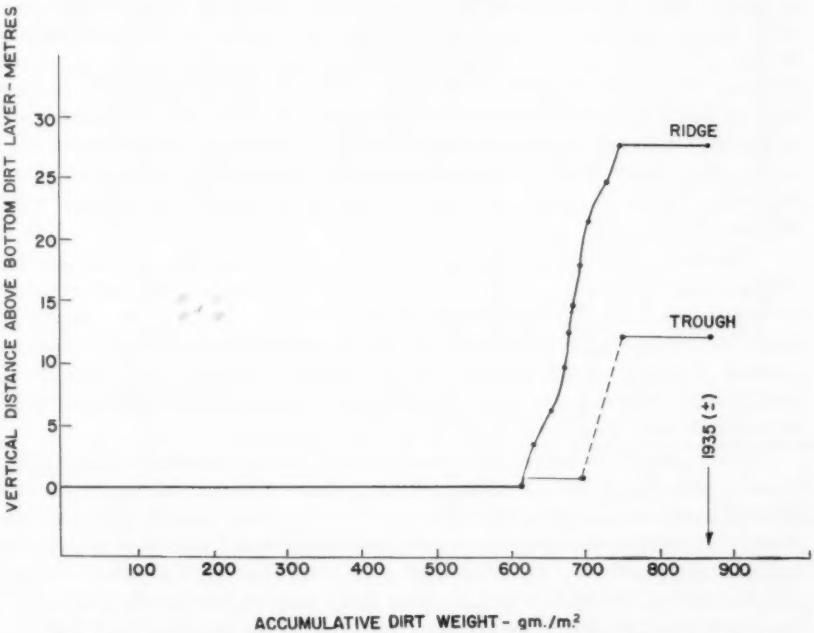


Fig. 23. Accumulative dirt weights under present trough and ridge on T-3.

Small magnetic spherules believed to be of meteoric origin were found by Stoiber *et al.* (1956) both in the surface dirt and in the lower dirt layers. They report: "The spherules show a range of size distribution. Of a group of 24 spherules isolated and measured . . . the diameter varied from 0.01 mm to 0.34 mm." Fifteen of the twenty-four measured 0.01 to 0.02 mm. "Smaller spherules were seen, but were difficult to identify with certainty and to isolate. . . . Characteristically, the micrometeoritic spherules are black, with either a striated or pitted surface. A few show

concave surface indentations analogous to those described in spherules from deep-sea cores. Very rarely they show a protuberance which appears to be a small bubble attached to the smaller sphere. A few of them are hollow. . . . Several methods were used to determine the composition of the magnetic spherules. All are magnetic. Some dissolve in aqua regia. Others do not. All dissolve in hydrobromic acid. They yield a positive microchemical test for iron, and a few of them gave questionable reactions for nickel. Although several attempts at spectrochemical analysis were made, no positive results were obtained because of the small amount of available material. On the other hand, X-ray powder analysis indicates that magnetite is a common constituent of many of the spherules and that olivine is present in some. Optical examination indicates that most of the spherules consist chiefly of a glass of mean refractive index of approximately 1.550 in which there is disseminated black opaque material (magnetite?) and some minor amounts of birefracting material (olivine?). Silicate glasses of this refractive index correspond to rocks of andesitic or basaltic composition with silica content of 50-60%. Our conclusion therefore is that most of the T-3 spherules which we have examined consist of silicate glass of the approximate chemical composition of a basalt or andesite and that in the glass there is disseminated magnetite and olivine. Compared with the large classes of meteorites, the T-3 spherules would be grouped with the aerolites."

"Several negative conclusions are of interest. The siliceous meteoritic tektites have glasses of lower refractive index than the T-3 spherulites. No spherulite from T-3 was found to be metallic, nor have we identified either native iron or iron carbides in the X-ray patterns. We have had no positive chemical determination of any element except iron. Nickel is questionable. Many tests were inconclusive because of the small amounts of material with which we were forced to experiment."

In an attempt to determine the age of T-3 by the number of spherules found in the dirt of the ice cores, a brief review of recent literature on the rates of infall was undertaken. This showed "extreme discrepancies in the present estimates, the higher values exceeding the lower by factors of approximately 500,000. The fact that meteoritic showers vary in intensity with locality on the earth's surface, that there may be considerable unrecognized industrial contaminant in many of the samples counted, and that the methods used to recover the spherules vary enormously in efficiency, makes some of the discrepancies understandable." However, from an analysis of these various reports, Stoiber *et al.* are of the opinion that the spherules found on T-3 are "far less abundant than one should expect had they been accumulating over a period of several thousand years." Obviously the uncertainties concerning the rate of accumulation and the percentage of recovery prevent any definite conclusions.

Salinity. Observations were made of the salinity of many of the lower sections of the ice cores, particularly those below the heaviest dirt layer. This was done by chloride titration methods with results shown in Fig. 24.

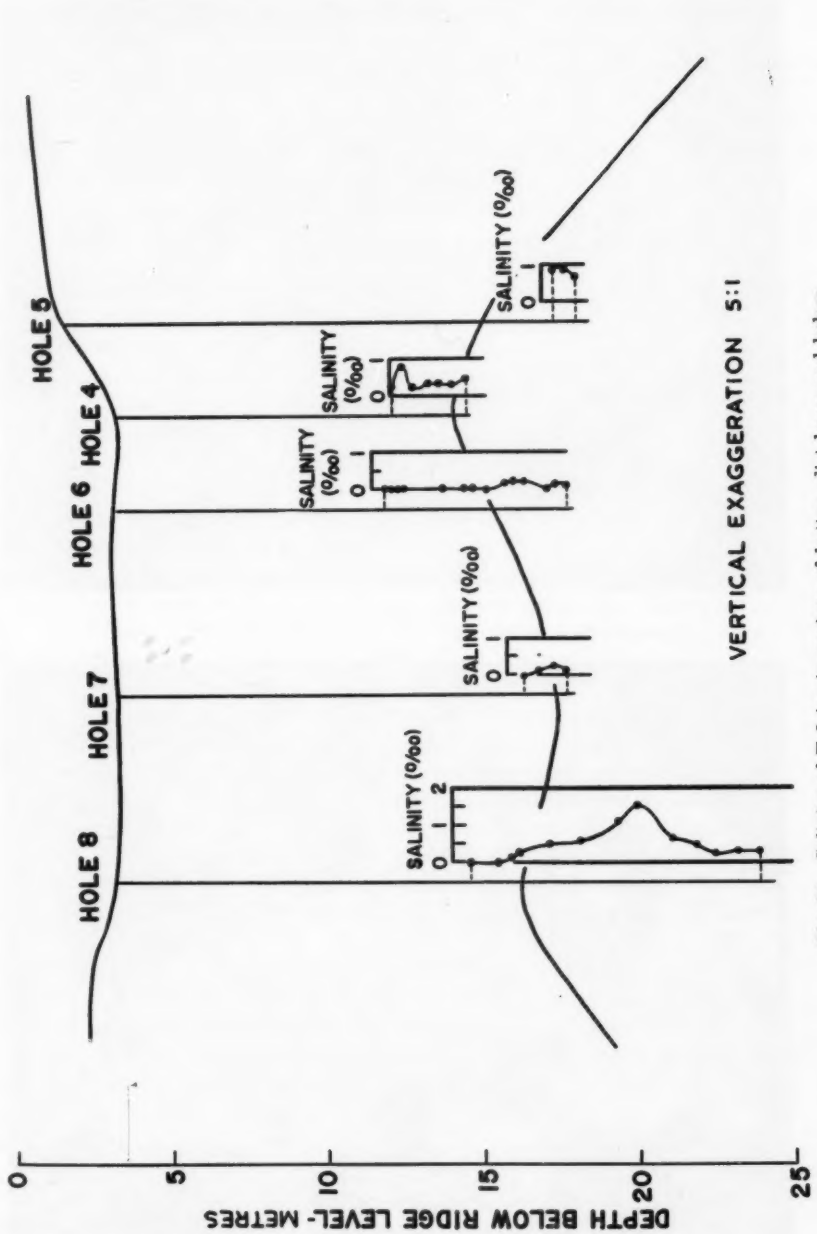
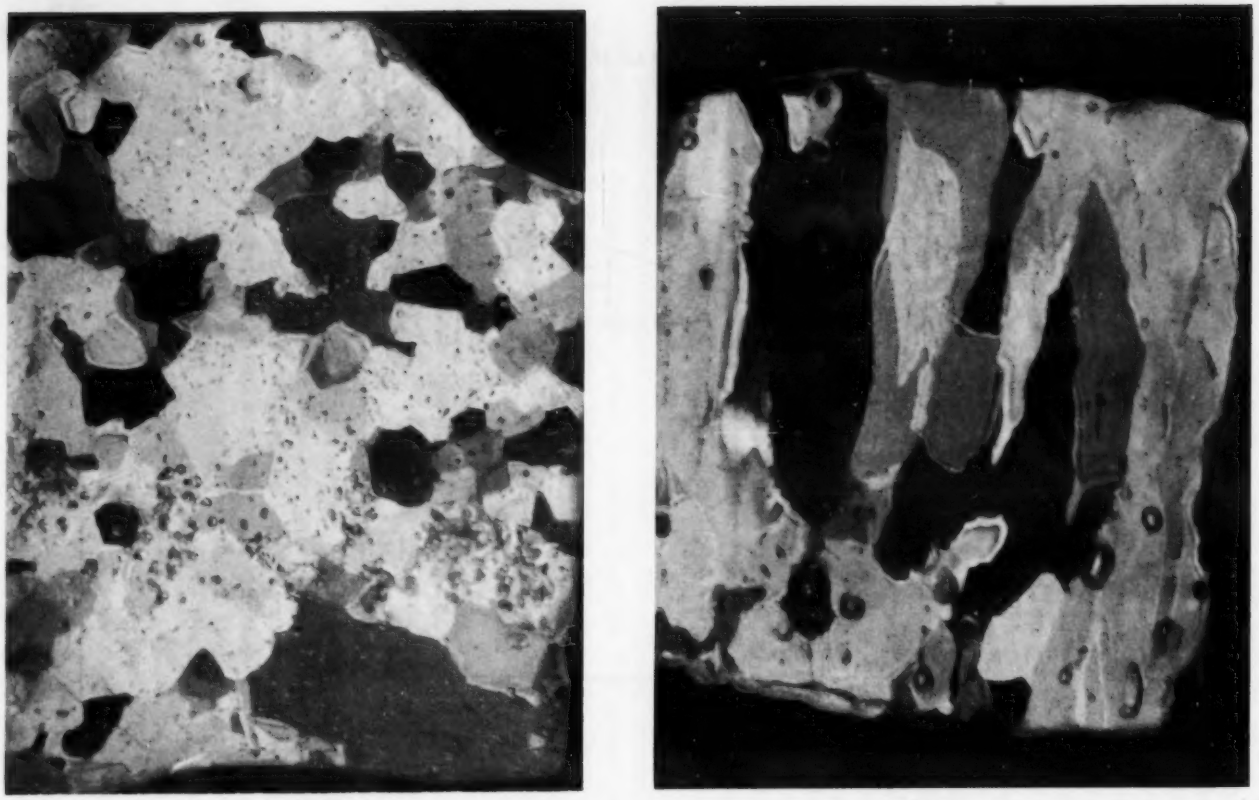


Fig. 24. Salinity of T-3 ice in regions of bottom dirt layer and below.



ICED FIRN LAKE ICE

Fig. 25. Typical thin sections of T-3 ice cores (magnification appr. X 1.2).

The salinities were all very low and usually zero above the heavy dirt layer, while only one observation of zero salinity was made below the heavy dirt layers. It would appear both from the dirt-free character, and the salinity, that the ice below the heavy dirt layer was ice growth accumulated from the bottom but, as noted later, observations of crystal character do not completely confirm this.

In the first deep hole drilled on the Ellesmere Ice Shelf in 1954 by Marshall, salt water encroachment occurred at about 24.5 metres and in a few days about 1 metre of salt water stood in the hole. This was also encountered on T-3 in Holes 4 and 6 near the top of the buried ridge. Here the damp ice persisted only for a fraction of a metre and then dry ice was encountered below. However, within a week about 2 metres of salt water had collected, which showed a salinity value of 54 parts per 1000, corresponding to the temperature of about —3°C prevalent at that depth. This salt water near the heavy dirt layer may be due to previous melting of salt ice layers, or it may have been collected from salt water which actually washed over the ice at an early date.

Crystal structure. On the basis of petrographic studies made in 1953 on the ice island and in 1954 on the Ellesmere Ice Shelf, Marshall (1955) has identified the upper ice as belonging to four types: iced firn, glacier ice, lake ice, and sea ice.

During the ice island program in 1955, photographs of thin sections of many of the ice cores were taken through crossed polaroids. Examples of these are shown in Fig. 25. W. F. Weeks (personal communication) has examined these data and has made a tentative separation of the ice into two types: (1) iced firn, which is semi-equigranular, has an average grain size of 1 centimetre and has regular, smooth, intergrain contacts and (2) lake ice, which shows a pronounced elongation of the crystals perpendicular to the freezing surface and more irregular, sutured, intergrain contacts in horizontal sections. None of the thin sections or rubbings were identified as either sea or glacier ice. It is possible that some of the material identified as lake ice could have originally been sea ice. This is considered doubtful however since Weeks was not able to observe any of the substructures so typical of sea ice. Weeks' identification of the ice types in the 1955 cores is shown in Fig. 26.

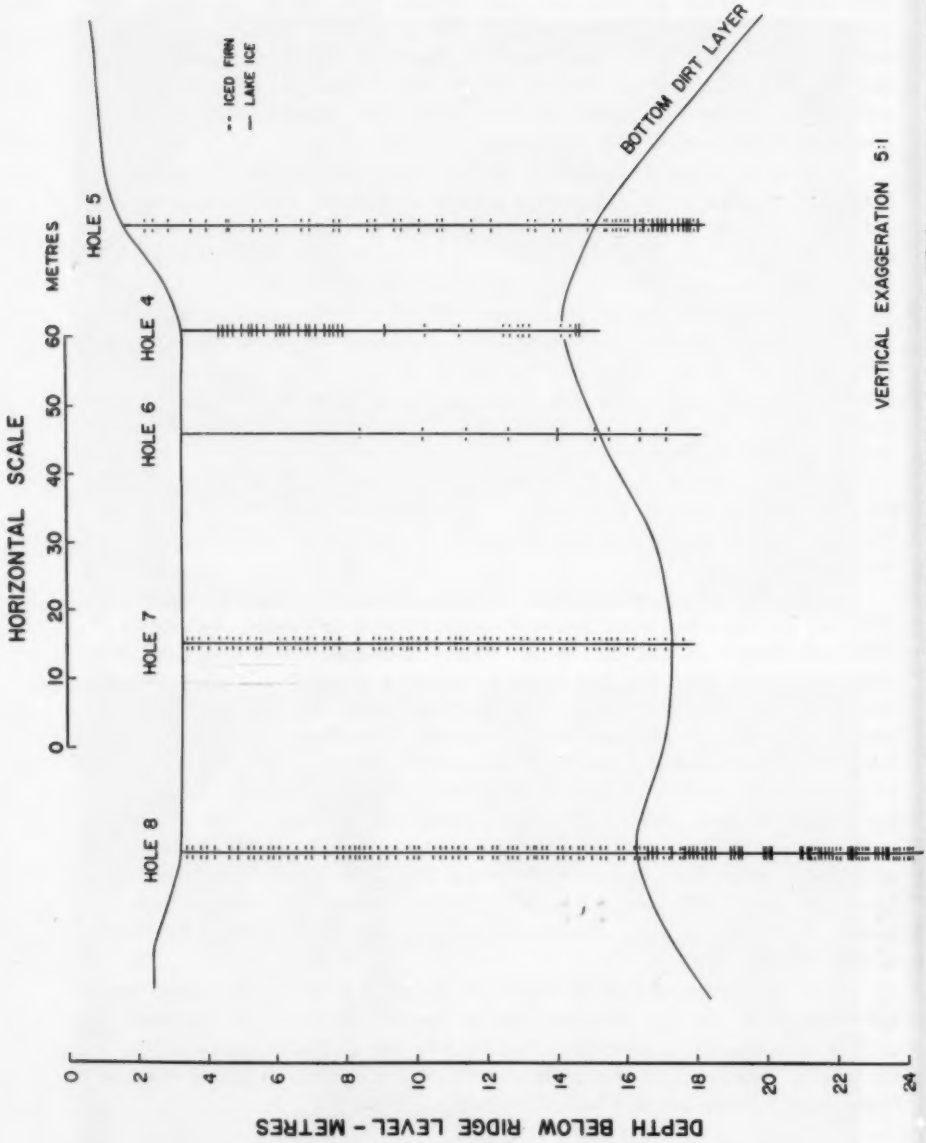
ICED FIRN

Fig. 25. Typical thin sections of T-3 ice cores (magnification appr. X 1.2).

It is interesting to note that the ice below the heavy dirt layer does not conform to the sea ice type, though usually showing the presence of salt. Further work is needed on this lower ice and also on the type of ice that is formed from brackish waters of low salinity such as may be found in the shore areas where the islands originated.

Internal ice temperatures

Copper-constantan thermocouples were located in six of the deep 8-cm-diameter holes drilled for petrographic and dirt layer studies; four of these were on the ice island, one on the Ellesmere Ice Shelf and one on the



Ward Hunt ice rise. These thermocouples were frozen in the holes and are still usable, serving also as long-term surface ablation indicators. Locations and depths below the ice surface at which thermocouples were originally installed are given in Table 2. Unfortunately the readings taken at T-3

Table 2. Locations and depths of thermocouples.

Location	*Depths (metres)*					*Date installed*
T-3 Ice Island, No. 1 - Ridge	3.0	9.1	15.2			Nov. 1952
T-3 Ice Island, No. 3 - Ridge	5.2	12.8	20.4	28.0		Sept. 1953
T-3 Ice Island, No. 7 - Trough	8.4	14.5				May 1955
T-3 Ice Island, No. 9 - Midway between Trough and Ridge	24.7					July 1955
Ellesmere Ice Shelf - Trough	0.3	1.5	3.0	4.6	6.1	
	7.6	9.1	12.2	15.2	21.3	May 1954
Ward Hunt Ice Rise	1.5	3.0	6.1	12.2	18.0	Sept. 1954

Hole 1 are not considered normal since very soon after the installation of the thermocouples, new camp buildings were erected near the site, with the consequence that abnormal drifting of snow occurred in the immediate area.

Shallow-depth temperatures. The best records of shallow-depth temperature variations were obtained in a ridge hole on the Ellesmere Ice Shelf between May and September 1954. They are shown in Fig. 27. Temperatures at greater depth were obtained at both Ellesmere and on the ice island and are shown in Fig. 28.

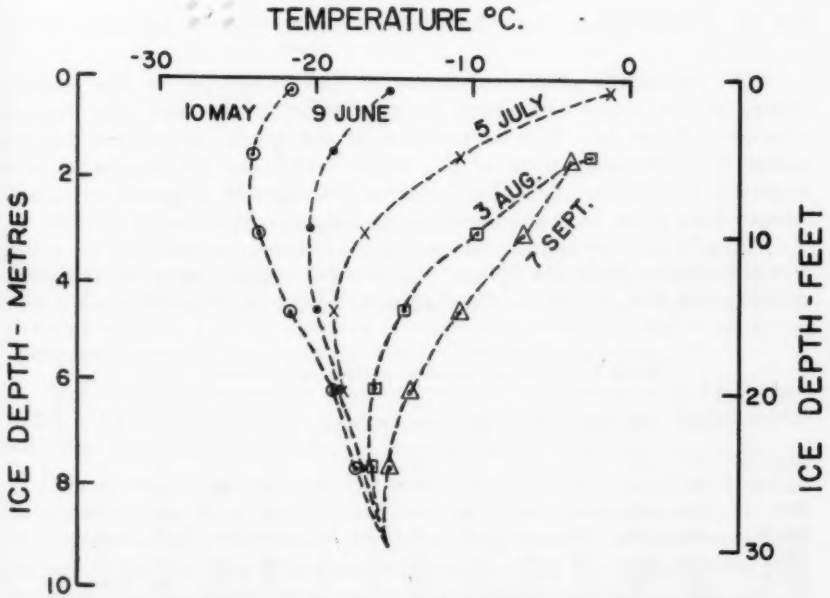


Fig. 27. Temperatures in a deep hole in the Ellesmere ice shelf, 1954.

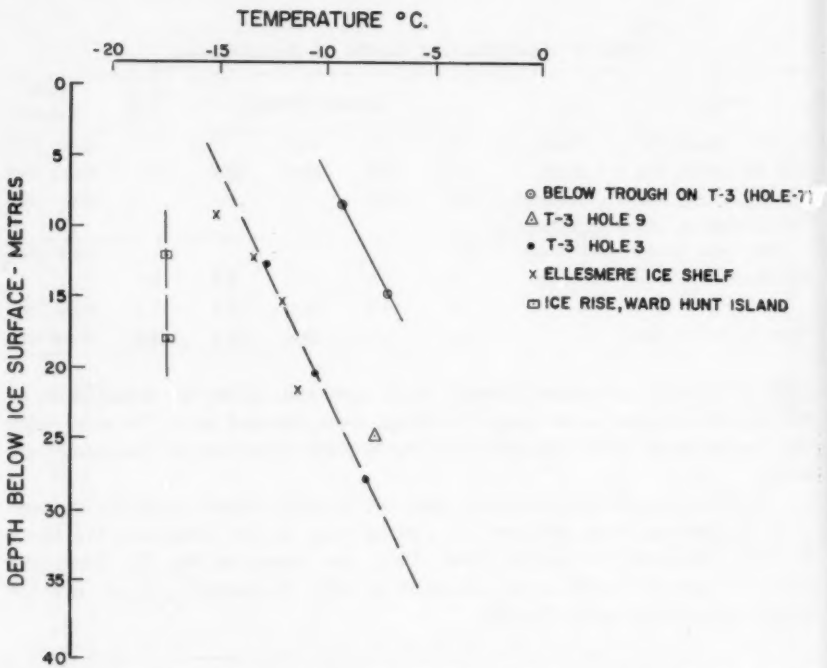


Fig. 28. Temperatures at depths below annual change, Ellesmere ice shelf and T-3.

The temperatures in the interior of the ice depend on the temperatures at the upper and lower boundaries of the ice and the thermal constants of the ice. The temperature of the lower ice surface may be taken as the freezing point of the water, which can be considered as a constant, though variation of as much as 2°C may be expected in coastal areas where there is an abundance of freshwater run-off under the ice.

Table 3 lists the monthly air temperature averages from the ice island T-3 and other high-latitude areas. The low T-3 values may be for abnormal years or areas but may also be explained by the very small upward flow

Table 3. Average monthly surface temperature (°C)

	Jan.	*Feb.*	*Mar.*	*Apr.*	*May*	*June*	*July*	*Aug.*	*Sept.*	*Oct.*	*Nov.*	*Dec.*	*No. of years recorded*
SEDOV	-28	-31	-31	-23	-12	-4	-2	-3	-8	-16	-21	—	1
FRAM	-36	-36	-31	-23	-11	-2	0	-2	-9	-22	-29	-32	3
T-3	-35	-37	-39	-24	-8	-2	0	-3	-11	-21	-27	-32	2
NP-2*	-37	-34	-28	-23	-10	-2	-1	-2	-8	-17	-28	-35	1
NP-3*	-34	-35	-29	-20	-11	-2	0	-2	-9	-14	-28	-30	1
NP-4*	-32	-37	-31	-26	-13	-2	0	-1	-9	-18	-31	-33	1

* Russian "North Pole" drift stations

of heat through the thick ice, compared to that of normal pack ice of 3 or 4 metres thickness. In Fig. 29 are shown average ice island air temperatures and most probable ice-surface temperatures. These latter are based on thermocouple or direct thermometer readings at various sites, but complete data are lacking, particularly in late winter months. It is noted that the curves are not sinusoidal, since the coldest and warmest months are March and July respectively. A theoretical fit to this type of curve is given by the equation:

$$T = A + B \sin \frac{2\pi t}{P} + C \sin \frac{4\pi t}{P}$$

where P = period, one year, t = time, T = mean monthly temperature, with the constant B about four times the numerical value of C. This curve is also shown in Fig. 29, where A = −17.75, B = 17.48, and C = 4.07.

The solution of the heat-flow equation in floating ice has been given by Malmgren (1927) for sinusoidal surface temperatures. It includes an additional term to the normal solution for a semi-infinite solid which is of equal value and opposite sign at the lower ice-surface boundary and decreases exponentially upward. This equation is given below.

$$T = T_1 + (T_2 - T_1)\frac{x}{l} + T'e^{-ax}\sin(nt - ax) - T'e^{-a(2l-x)}\sin[nt - a(2l-x)]$$

where T = mean monthly temperature
T_1 = mean annual temperature at the surface
T_2 = temperature at lower ice boundary
l = total ice thickness
x = depth
t = time
T′ = amplitude
$a = \sqrt{\frac{n}{2\alpha}}$
$n = \frac{2\pi}{P}$, P = period
α = thermal diffusivity

With thermal constants of freshwater ice, the standard solutions for a semi-infinite solid can be used at depths below 10 metres with the addition of a linear term. For the equation of arctic ice surface temperatures given above, each term of the equation can be handled separately.

In Fig. 30 are shown the observed values of temperature at 4.6 metres in the Ellesmere Ice Shelf compared with theoretical values. Small variations are to be expected in different areas and different years.

Temperatures below depth of annual change. Temperatures obtained at the ice rises near Ward Hunt Island of −17.7°C at 12.2 metres and −17.3°C at 18.0 metres (Fig. 28) should be the average annual temperature at the surface of the ice. The differences occurring during the course of the year between the monthly average temperature of the ice surface and that of the air, as recorded at the normal height of 2 metres at surface weather

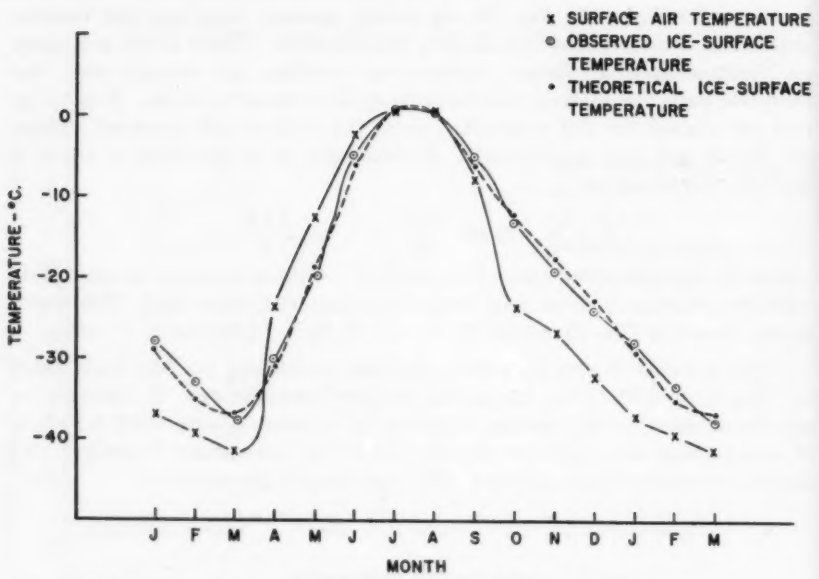


Fig. 29. Average ice island temperatures and most probable ice surface temperatures.

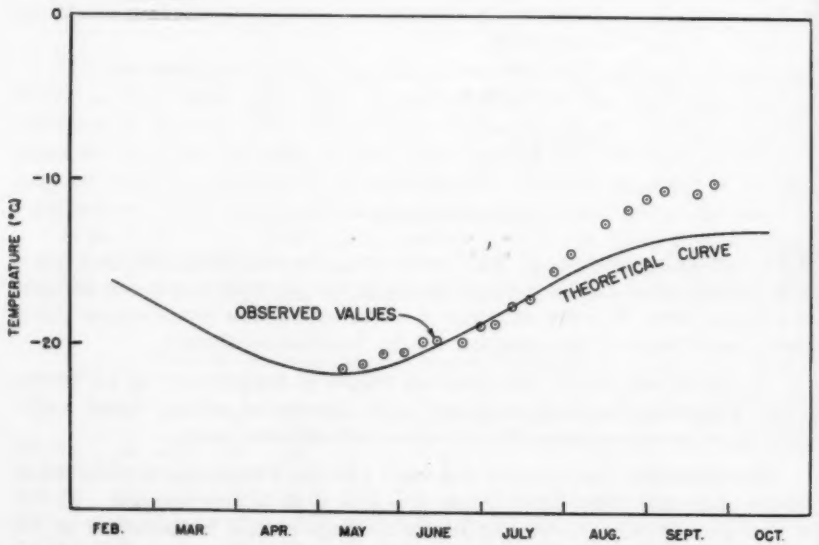


Fig. 30. Observed values of temperatures at 4.6 metres in the Ellesmere Ice Shelf compared with theoretical values.

stations, may not be very large, as is indicated in Fig. 29, despite the effects of snow cover and long-wave radiation from the ice surface. The annual average air temperatures at the weather stations Alert and Eureka, the two nearest to Ward Hunt Island, are —16.8°C and —19.1°C respectively, taken over a period of 3 years.

The linearity of the temperatures under the ridges, between surface average and the temperature of freezing water, shows that the ice shelf and island must consist of ice of apparently uniform thermal conductivity. Underneath the troughs temperatures are somewhat higher, as would be expected, considering that the drainage lakes remain after summer melting has ceased. At the bottom of the lakes of average depth the temperature would be 0°C for four to six months and a linear relation would be established between the resulting higher annual average and the sea water temperatures. Directly under the troughs, therefore, the gradient in the bottom part of the ice would be less than that under the ridges, which could tend to build up less ice there. This effect may be decreased somewhat in the areas near land if fresh melt-water is present in sufficient quantities to extend under the shelf. The lighter, less saline waters with higher freezing temperatures would be found under the shallow part of the shelf and this should tend to offset the influence of the smaller gradient.

In 1952 a freshwater lake under considerable pressure was located in the trough area adjacent to the campsite on T-3 (see Fig. 31). Two dirt layers were found in the 2.5 metres of ice above this lake, so that it was believed at least two melt seasons had occurred since the lake had formed.

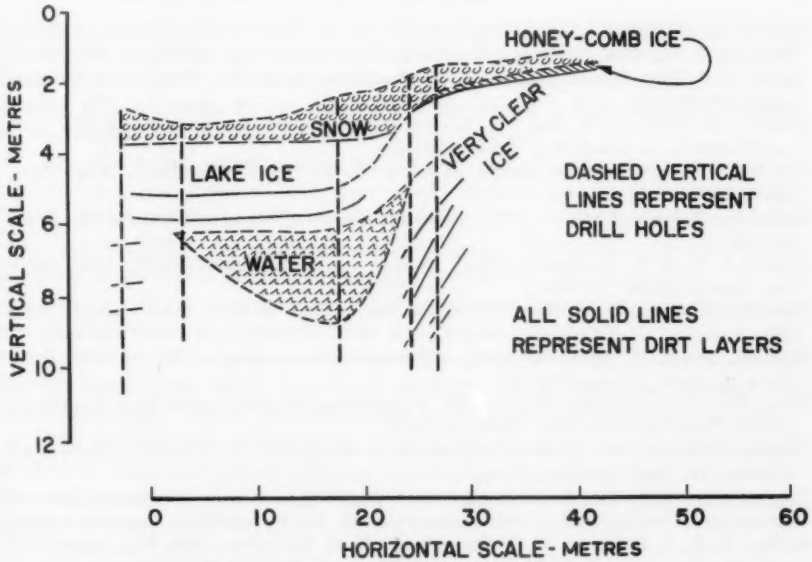


Fig. 31. Buried lake found on T-3 in 1952.

In 1953 at the same location there were 3 metres of ice over this lake. It was also noticed that clear ice with very large crystals was characteristic of the surroundings of this lake, presumably caused by the high pressures due to expansion during freezing. In 1955 (see Fig. 22) this lake was entirely frozen. It is presumed that this lake was formed in the summer of 1949 or 1950 when the ice island was at the most southerly part of its track.

Acknowledgements

The author wishes to express his appreciation for the assistance of the following personnel of the Geophysics Research Directorate, Air Force Cambridge Research Center: R. D. Cotell, T. F. Sexton, R. R. Shorey, and Norman Goldstein for many of the field operations on the ice island T-3. Renner Hofman, for study and interpretation of some of the seismic records. Wilford Weeks for study of ice photos and etchings. Vivian Bushnell for data reduction and preparation of illustrations.

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The many discussions with E. W. Marshall of the Snow, Ice, and Permafrost Research Establishment and Dr. G. Hattersley-Smith of the Defence Research Board of Canada were most helpful.

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GLACIATION ON THE ARCTIC SLOPE OF THE BROOKS RANGE, NORTHERN ALASKA*

R. L. Detterman, A. L. Bowsher, and J. Thomas Dutro, Jr.†

THE area discussed in this report is the Arctic Slope of the Brooks Range, northern Alaska. The Brooks Range extends eastward across northern Alaska from Cape Lisburne to the Canadian border (Fig. 1). Its crest lies about 95 miles north of the Arctic Circle and approximately parallel to it. Near the eastern end the peaks rise to 8,000 or 9,000 feet above sea-level. In the central part of the range the higher summits attain about 7,000 feet. The mountains become progressively lower westward toward the headwaters of the Noatak River. At Howard Pass, approximately 2,200 feet above sea-level, the surrounding mountains are about 5,000 feet high. The DeLong and Baird Mountains, westerly extensions of the Brooks Range on either side of the broad valley of the west-flowing Noatak River, rise to between 4,000 and 5,000 feet above sea-level.

The streams on the north side of the range flow generally northward through deep, U-shaped, glacially sculptured valleys that traverse the Arctic Foothills Province of rolling hills and east-west trending ridges. In the eastern part of the region the streams flow directly into the Arctic Ocean, but in the western part they run first into the Colville River, which flows east and north into the sea. A low ridge that trends east-west lies north of the Colville River. To the north of this ridge extends the Arctic Coastal Plain Province to the shores of the Arctic Ocean.

Most existing glaciers on the northern slope of the Brooks Range originate in the protected parts of the larger, higher, north-facing cirques and do not extend more than 3 or 4 miles from the areas of accumulation. The smaller glaciers seem to be remnants of shrinking valley glaciers. Almost all the smaller cirques at lower levels are ice-free.

Six glaciations of the area have been recognized by an analysis of morphology, composition, and geographic distribution of glacial deposits. Glaciers were of alpine valley type that during the earlier, more extensive advances coalesced to form piedmont lobes. There was no continuous ice-sheet in northern Alaska during any part of the glacial epoch.

The two oldest advances, the Anaktuvuk and Sagavanirktok glaciations, are characterized by deposits that are greatly modified by erosion and mass-wasting. Drift is present as tundra-covered moraines. Most of the former kettle lakes have been filled or drained and drainage patterns are well integrated. These glaciations are considered pre-Wisconsin in age.

The Itkillik and Echooka glaciations, represented by relatively fresh drift, are distinguished by differences of physical characteristics and the distribution of morainal areas. They are thought to be of early Wisconsin age.

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* Publication authorized by the Director of the U. S. Geological Survey.

† U. S. Geological Survey.

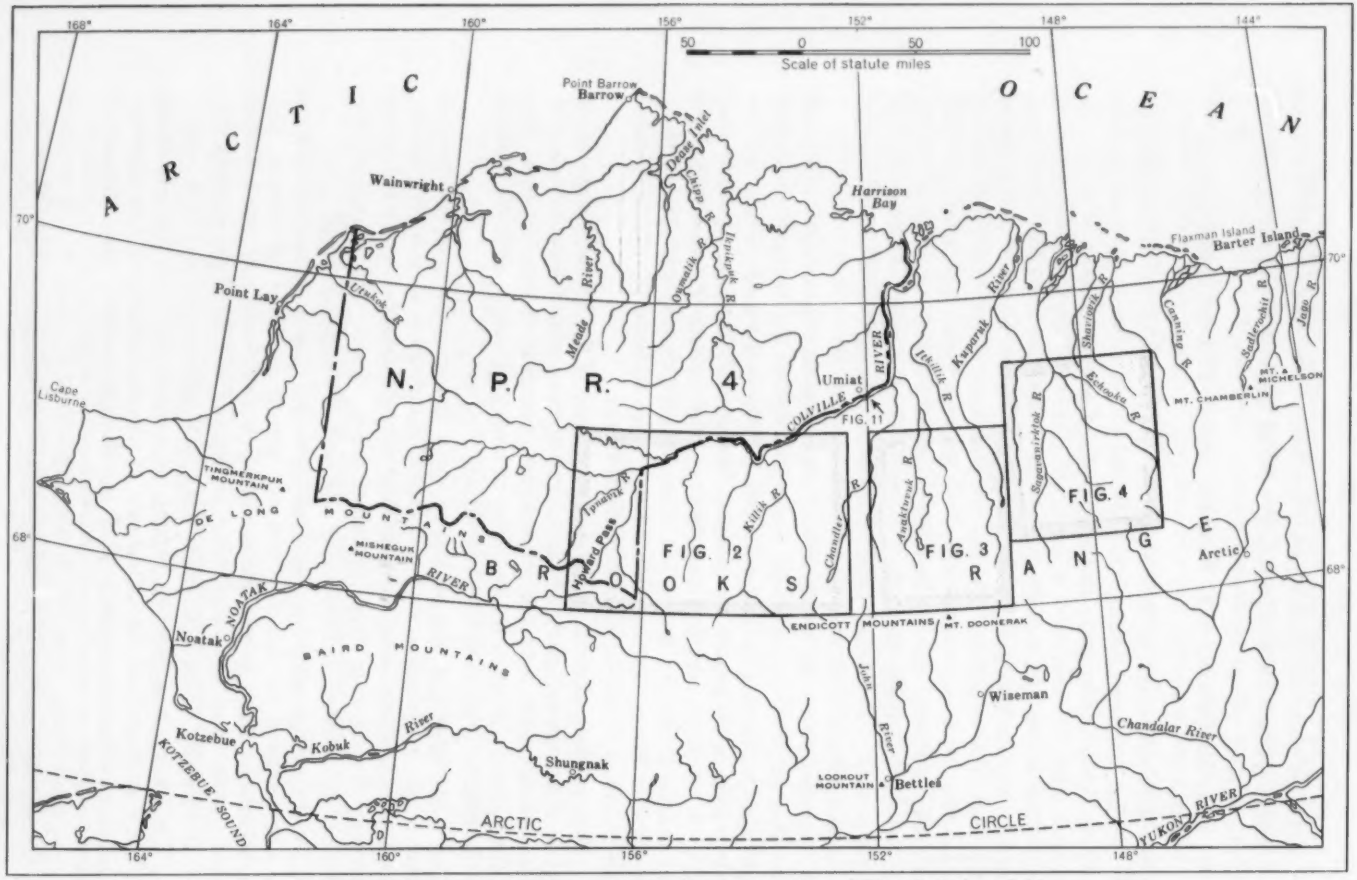

Fig. 1. Index map of northern Alaska showing location of Figs. 2, 3, 4, and 11.

Deposits of the Alapah Mountain glaciation are little modified by weathering, but are tundra-covered. Morainal areas contain kettle lakes and consequent drainage has undergone little or no modification. This advance is tentatively dated as late Wisconsin.

The Fan Mountain glaciation represents the most recent glacial advance recorded in the region. Moraines are fresh, bare of tundra and generally restricted to the cirque areas. It was obviously a very recent event.

Outwash deposits and loess extend beyond the limits of the northernmost moraines, nearly to the arctic coast. The Gubik formation (Gryc, Patton, and Payne, 1951, p. 167)[1] of the Coastal Plain area is overlain by a terrace deposit that can be traced to the margin of the Itkillik till sheet. Outwash deposits of the two older glaciations may possibly be included in the Gubik formation as it is now mapped.

Fig. 1. Index map of northern Alaska showing location of Figs. 2, 3, 4, and 11.

Previous work

F. C. Schrader (1902, 1904) was the first geologist to report on glacial phenomena observed in field studies of the Brooks Range. These papers, which contain descriptions of features observed on a reconnaissance up the John River and down the Anaktuvuk River to the Arctic Ocean, present excellent discussions of the glacial deposits that characterize the Brooks Range throughout its length. E. deK. Leffingwell (1919, p. 133) presented detailed discussions of glacial features in the Canning River region in the eastern part of the Brooks Range. P. S. Smith and J. B. Mertie, Jr. (1930, p. 242-247) summarized the knowledge of glaciation in northwestern Alaska and presented data collected during their work on the Arctic Slope and adjoining areas. A few summary remarks about extent and type of glaciation in the Brooks Range are included in papers by S. R. Capps (1931) and P. S. Smith (1939, p. 67-70).

Detterman (1953, p. 11-12) discussed glacial deposits in the Sagavanirktok-Anaktuvuk region, northern Alaska, and proposed the names Anaktuvuk, Sagavanirktok, Itkillik, and Echooka glaciations for the four Quaternary glacial advances recognized in that region.

Present investigations

This report is intended to provide a more detailed account of the glacial history of northern Alaska than was given by Detterman (Péwé *et al*, 1953) and is based on observations made during investigations of the Naval Petroleum Reserve No. 4 by the U. S. Geological Survey during the period 1944 to 1953 of the nature and extent of glacial deposits. Field parties followed most of the larger rivers on the Arctic Slope, travelled over much

[1] The Gubik formation of Pleistocene age ranges in thickness from a few to 150 feet, but is 10 to 30 feet thick in most exposures. It consists predominantly of slightly consolidated, brown or buff gravel, sand, silt and clay.

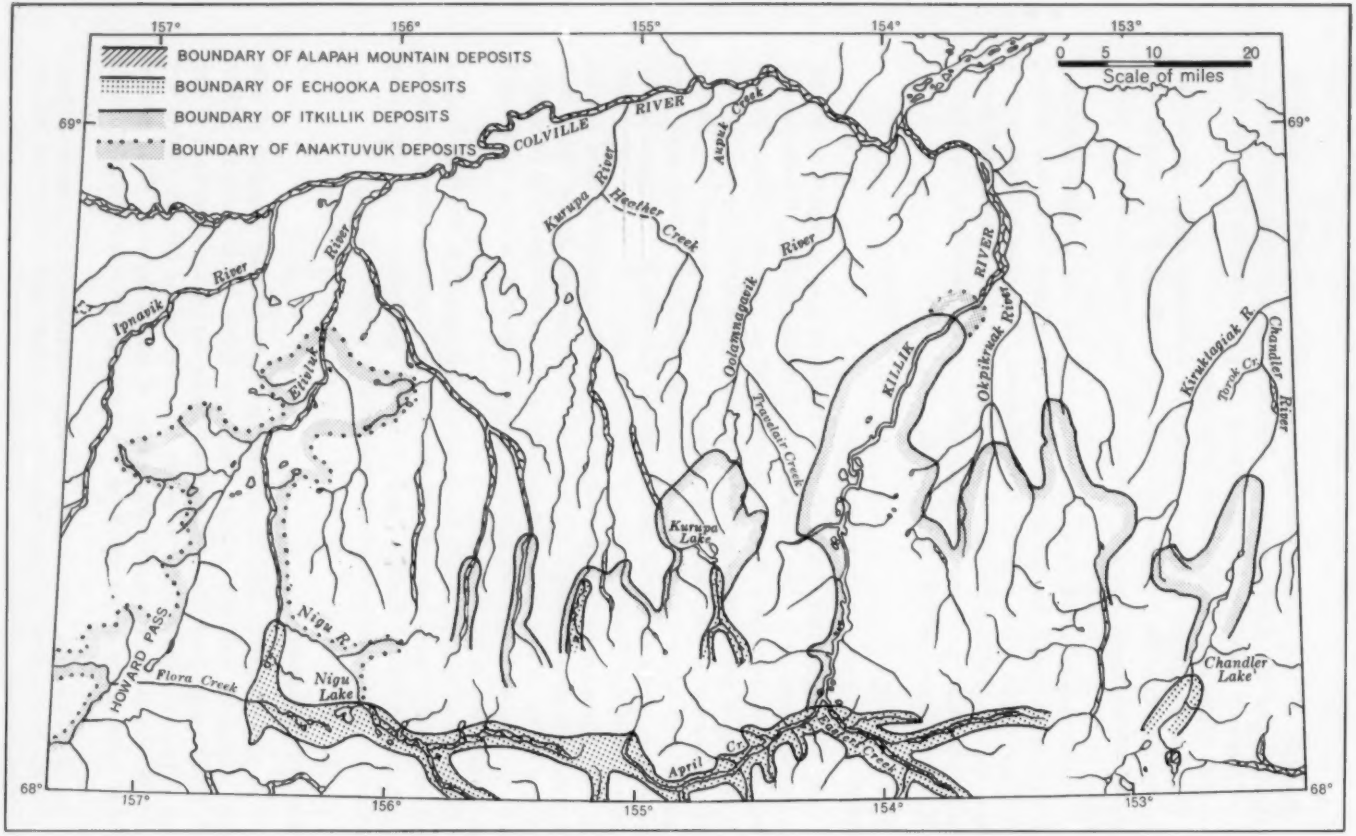

Fig. 2. Distribution of glacial deposits in the Ipnavik-Chandler rivers region.

of the Foothills Province in tracked vehicles and studied many local areas supported by aircraft equipped with ski or floats. Generally the glacial studies were incidental to work on structure and stratigraphy. Nevertheless, much information on glaciation was obtained for a considerable part of northern Alaska.

This report and the maps showing the distribution of glacial deposits are largely based on work by R. L. Detterman, M. D. Mangus, and R. M. Chapman along the Killik River (Fig. 2), by A. L. Bowsher and J. T. Dutro, Jr., around the Shainin, Nanushuk, and Itkillik lakes (Fig. 3), and by R. L. Detterman and A. S. Keller in the eastern part of the Foothills Province (Fig. 4).

In addition, several other geologists of the Geological Survey have contributed remarks and supplementary data on glaciation. Work in the region has been so intimately interwoven that it is not possible to acknowledge every contribution.

This report is concerned with glacial deposits found in the central part of the Brooks Range between Shainin and Itkillik lakes and in the southern part of the Foothills Province from the Shaviovik River in the east to the Etivluk River in the west. Remarks herein concern glaciation only as it affected the north side of the Brooks Range. None should be construed as applying to glaciation on the south side of the mountain chain.

Glacial deposits

Anaktuvuk glaciation

The Anaktuvuk glaciation (Detterman, 1953) is named for fragmentary morainal remnants on the Anaktuvuk River, 40 miles north of the mountain front (Fig. 3). The southern edge of the deposit is about 25 miles north of the mountains; it is obscured locally by outwash gravel and till of a younger stage of glaciation.

Deposits of this glaciation have been identified on the Itkillik, Nanushuk, Killik, Nigu, and Etivluk rivers and are probably present on most of the other streams, as this glaciation was apparently more extensive than any of the succeeding advances. Till in the inter-stream areas and erratics on the ridge tops indicate that valley glaciers coalesced to form a piedmont lobe. Conglomeratic quartzite boulders are present on the Nanushuk and Killik rivers 10 to 15 miles north of the morainal belt and may indicate that glaciation was more extensive than mapped on Figs. 2 and 3. On the other hand, these erratics may represent a glaciation that pre-dates the Anaktuvuk advance.

The lower peaks in the Brooks Range, as well as the majority of the ridges in the foothills adjacent to the mountains, were probably covered with ice during the glacial maximum. That the source of some of the ice was south of the mapped area is indicated by granite boulders in the Howard Pass area; the nearest granite is in the mountains along the

Fig. 2. Distribution of glacial deposits in the Ipnavik-Chandler rivers region.

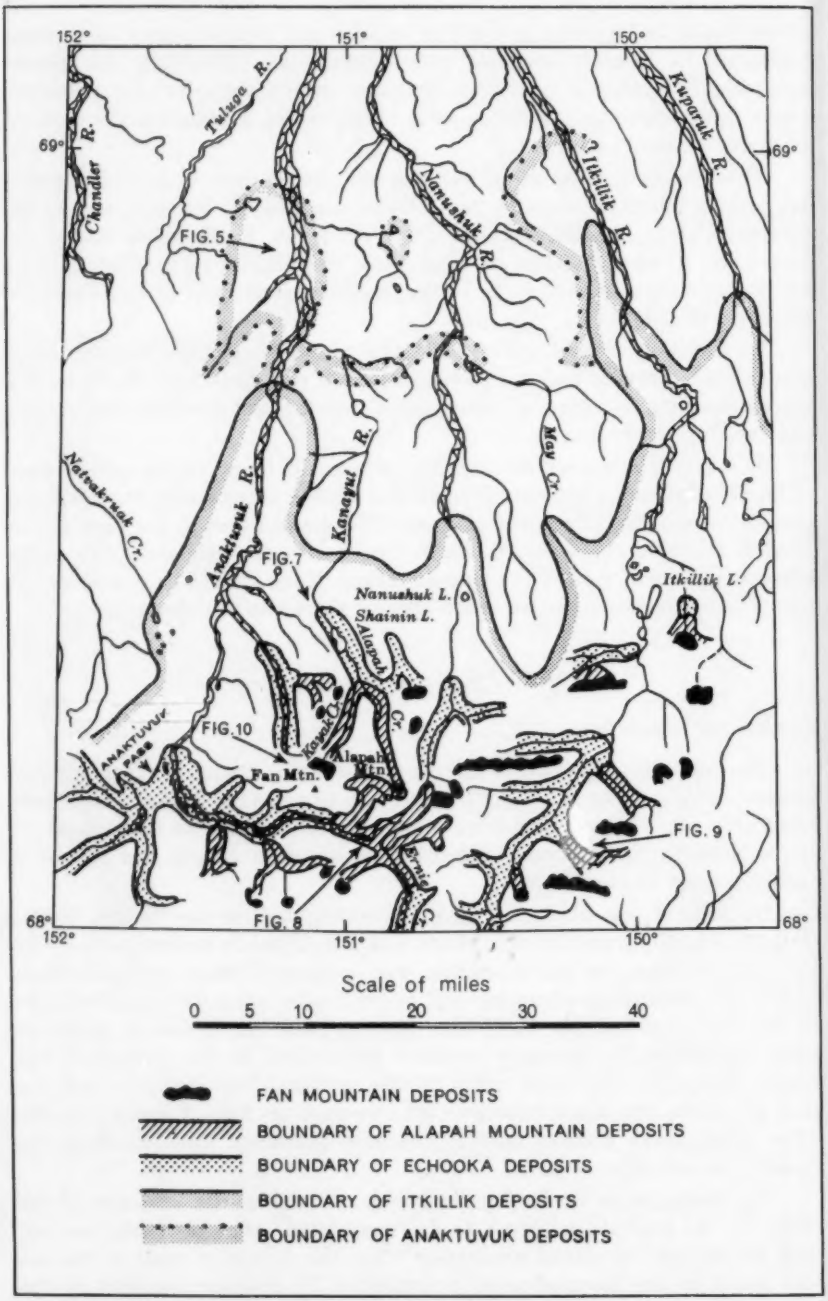

Fig. 3. Distribution of glacial deposits in the Anaktuvuk-Kuparuk rivers region.

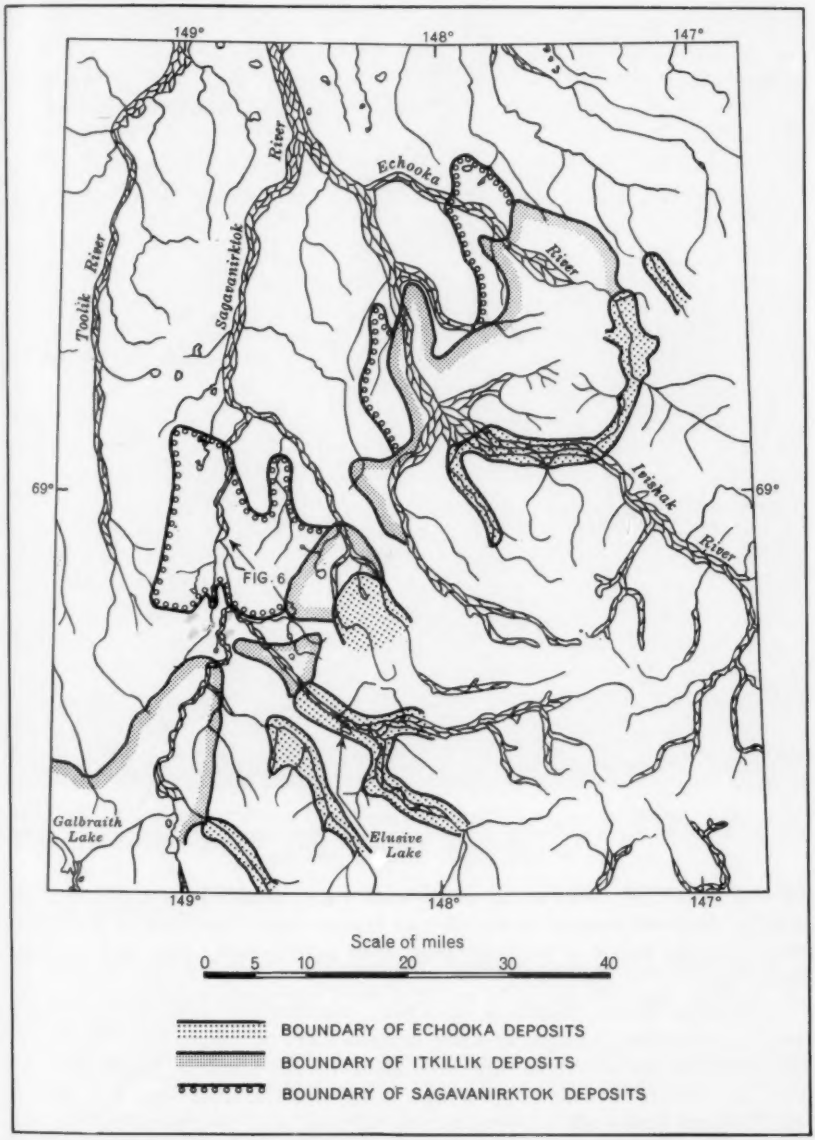

Fig. 4. Distribution of glacial deposits in the Sagavanirktok-Echooka rivers region.

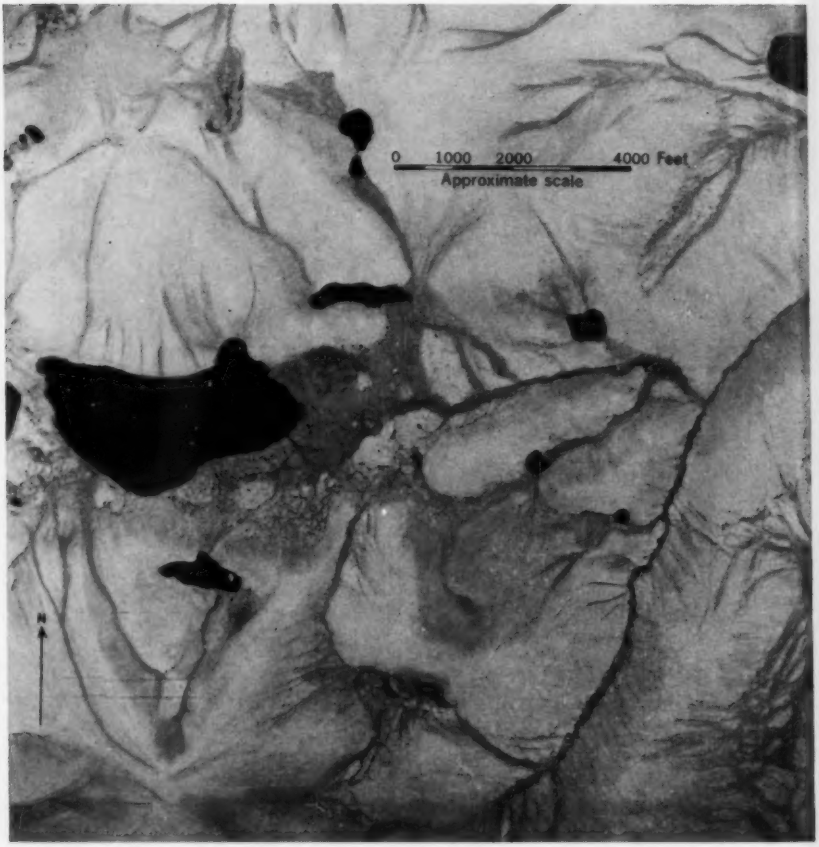

Photo: U. S. Navy

Fig. 5. Drift of the Anaktuvuk glaciation on the west side of the Anaktuvuk River, 30 miles north of the Brooks Range.

Noatak River, 30 miles south of Howard Pass. Glacially sculptured topographic features present in the Brooks Range today may date in part from this advance, but it is believed they were considerably modified by later advances.

Deposits of this glaciation have been considerably modified by erosion and mass-wasting. Lateral and end moraines are no longer identifiable and the drift is preserved as low, rounded, tundra-covered, morainal hills (Fig. 5). Large erratics of quartzite, conglomeratic quartzite, and granite (in Howard Pass) are present on the hilltops, and the morainal deposits consist of smaller fragments of similar composition; any fragments of shale, sandstone, or limestone that may have been present on or near the surface have since disintegrated. Most of the kettle lakes have been filled or drained and the streams are well integrated.

Sagavanirktok glaciation

The Sagavanirktok glaciation (Detterman, 1953) is named for morainal remnants along the Sagavanirktok River (Fig. 4), 60 miles east of the Anaktuvuk River. The deposits are separated from younger ones to the south by outwash gravel and their northern limit is about 25 miles north of the Brooks Range.

Deposits of this glaciation have been recognized only in the eastern part of the mapped area, along the Sagavanirktok, Ivishak, and Echooka[2] rivers; they have not been recognized in areas where deposits of the Anaktuvuk glaciation are present, nor have deposits of the Anaktuvuk glaciation been recognized in areas where the Sagavanirktok glaciation is mapped. The exact relationship between these two advances is not known and it is possible that the Sagavanirktok deposits represent a readvance during the Anaktuvuk glaciation or they may actually be deposits of the Anaktuvuk glaciation. The difference in superficial modification of the deposits, however, is great enough to permit distinguishing them and it seems to indicate a time gap between the two glaciations.

Morainal remnants of the Sagavanirktok glaciation have been considerably modified, but still retain recognizable knob and kettle topography (Fig. 6). Many of the kettle lakes have been filled and streams are well integrated. The lateral and terminal moraines are recognizable in a few places, but for the most part are covered by outwash of more recent advances. Most of the erratics and morainal debris consist of quartzite and conglomeratic quartzite similar to the deposits of the Anaktuvuk glaciation. Limestone fragments are present and may locally constitute as much as 30 per cent of the drift; they are rarely found in the deposits of the Anaktuvuk glaciation.

Deposits along the Sagavanirktok River indicate that the ice coalesced locally to form a piedmont lobe in the foothills area. The depth of ice that filled the mountain valleys is unknown, but quartzite boulders 1,500 feet above the valley floor may date from this glaciation.

Itkillik glaciation

The Itkillik glaciation (Detterman, 1953) is named for deposits that are well developed along the Itkillik River (Fig. 3), 30 miles east of the Anaktuvuk River.

Deposits of this glaciation are present on all the major streams in the mapped area. The northernmost extent of the advance was about 25 miles from the mountains along the Itkillik River, 10 miles from the mountains in the eastern part of the mapped area, and 30 miles from the mountains along the Killik River.

[2] The spelling "Eschooka" instead of "Echooka" in Table 1 of U. S. Geol. Surv. Circ. 289, "Multiple Glaciation in Alaska," is a typographical error.

This glaciation was the last advance of ice that extended north beyond the mountain front. Piedmont lobes formed in the foothills near the major river valleys; all subsequent advances were restricted to valley glaciers within the mountains. Most of the ice-sculptured topographic features in the mountains probably date from this glaciation. In the Killik River valley the upper limit of ice-action is about 1,800 to 2,000 feet above the valley floor and erratics are found 1,200 feet above the river in the foothills north of the mountains.

Deposits of the Itkillik glaciation are relatively fresh and only slightly modified by erosion. Lateral and terminal moraines are easily recognized, as well as knob and kettle topography of the ground moraine (Fig. 7). Kame terraces and ice-margin stream-channels are present along the valley walls of the larger rivers; numerous kettle lakes, a few tens of feet to half

Photo: U.S.A.F.

Fig. 6 Deposits of the Sagavanirktok glaciation (background) and Itkillik glaciation (foreground) on the Sagavanirktok River, 12 miles northwest of Elusive Lake.

Photo: U.S.A.F.

Fig. 7. Moraines of the Itkillik glaciation (foreground) on the east side of Anaktuvuk River and moraines of the Echooka glaciation at the north end of Shainin Lake (right background). Shainin Lake is dammed by Echooka moraines.

a mile in length, are present on the moraines. The small streams are not well integrated and drainage patterns are arcuate in outline around the perimeter of the moraines. Most of the large lakes in the river valleys at the northern front of the mountains were formed through damming of streams by deposits of this glaciation. Six recessional moraines are present in the Killik Valley, one of which dammed a lake 29 miles long. This lake was partly drained when the river cut through the moraine.

Erratics of the Itkillik glaciation have well developed glacial striations. Locally limestone forms as much as 50 per cent of the erratics and larger rock fragments in the moraines. It appears to be relatively fresh and is only slightly weathered, in contrast to the highly weathered limestone in the older deposits.

Echooka glaciation

The Echooka glaciation (Detterman, 1953) is named for the morainal deposits along the Echooka River, a tributary of the Sagavanirktok River in the eastern part of the mapped area (Fig. 4). This relatively minor advance did not reach the mountain front and was confined to the larger river valleys. Small tributary glaciers probably coalesced to form trunk glaciers in the larger valleys, but piedmont lobes were not formed. The more extensive deposits of this glaciation are in the headwaters area of the Killik and Nigu rivers. There are other large areas on the upper parts of the Anaktuvuk, Itkillik, and Kanayut rivers (Fig. 7). Similar, unmapped deposits are probably present in other river valleys.

The Echooka glaciation left well developed, arcuate terminal moraines that cross the river valleys near the mountain front. The kame terraces along the valley walls and the knob and kettle topography of the ground moraines are quite fresh. Kettle lakes are small and have not been filled or drained. Small consequent streams crossing the moraines have not cut deeply into them. The larger rivers, however, have removed a considerable amount of the morainal material from their flood plains.

Drift of this glaciation forms a thin mantle 30 to 50 feet thick, although the lateral and terminal moraines are somewhat thicker. Rock fragments consist of fresh-appearing limestone, sandstone, and quartzite, which are all striated and have fresh, ice-polished surfaces. Small rock fragments are common. The average diameter of erratics is about 18 to 24 inches, whereas most of the drift is composed of gravel and cobbles as much as 6 inches in diameter.

End moraines of the Echooka glaciation are about 20 to 40 miles from the accumulation areas in the cirques. The maximum ice-movement was on the Killik and Nigu rivers, where end moraines are as far as 50 miles from the cirques. Shainin and Elusive lakes were formed behind end moraines of this advance (Fig. 7).

End moraines of the Echooka glaciation are present in nearly all stream valleys in the Brooks Range. The occurrence of end moraines at about the same position in relation to the mountain front is one of the features distinguishing deposits of the Echooka glaciation.

Alapah Mountain glaciation

The Alapah Mountain glaciation is here deduced from morainal remnants found in the pass between Anaktuvuk River and Ernie Creek (Fig. 8), 16 miles south of Shainin Lake (Fig. 3), these deposits having been formed at the snouts of the glaciers originating on the slopes of Alapah Mountain.

Deposits of this glaciation are not extensive, but small remnants are present throughout the Brooks Range. They have been washed from many of the glacial valleys; where present they lie well back in the range.

Photo: U.S.A.F.

Fig. 8. Alapah Mountain end moraines in the Anaktuvuk River-Ernie Creek Pass, 4.5 miles southwest of Alapah Mountain.

Poorly developed deposits are recognized 4 miles southeast of Itkillik Lake; reduced remnants of till occur 5 miles southwest of Shainin Lake. Well defined remnants are present at the junction of Alapah and Kayak creeks, 4 miles south of Shainin Lake and there are other deposits in tributary

Photo: U.S.A.F.

Fig. 9. U-shaped glacial valleys and glaciers at the head of the Itkillik River, 22 miles south of Itkillik Lake. Fan Mountain deposits are present in some of the cirques. Alapah Mountain deposits are present along walls of the higher valleys and lie along the side of the main valley in the centre of the photograph.

valleys at the head of Anaktuvuk River. Extensive lateral moraines occur in tributaries at the head of the Itkillik River (Fig. 9).

Terminal moraines of the Alapah Mountain glaciation are differentiated from those of the Echooka glaciation partly by their geographical position; they lie 5 miles or more upstream from the Echooka terminal moraines. Moraines of the Echooka glaciation are more extensive, thicker, and slightly more modified by mass-wasting and erosion than those of the Alapah Mountain glaciation. Generally more extensive drift deposits are present between the Echooka and Alapah Mountain terminal moraines than upstream from the last named. The lower walls of the valleys upstream from the Alapah Mountain terminal moraines are characterized by steep slopes with cliffs cut into bedrock. Very little talus has accumulated at the

base of these slopes and cliffs. Downstream from the Alapah Mountain end moraines talus fans extend well out into the valleys; terraces and irregular patches of drift lie along the base of the valley walls.

Deposits of the Alapah Mountain glaciation, largely covered by tundra, produce conspicuous morainal ridges and patches of till with relatively smooth, rolling surfaces. Some small kettle lakes are present between morainal ridges. Most of these clear-water lakes with gravel-covered bottoms lie in poorly drained areas; very little silt or sand has been transported into them. Drainage is, in general, consequent and controlled by the position of the drift ridges. Drainage channels and the three or four morainal ridges commonly present clearly outline the margin of the ice from which the deposits were derived. The deposits seem to represent a single glacial stagnation. Upstream from the moraines small patches of drift are preserved as terraces or more rarely as lateral moraines along the sides of the valleys. Terminal moraines are found from 3 to 10 miles downstream from the cirques where the ice accumulated.

The deposits of the Alapah Mountain glaciation consist largely of remnants of unsorted drift, composed of limestone, sandstone, and shale fragments derived from nearby cirques.

Short valley glaciers characterized this glaciation, which represents a late, short-lived advance.

Fan Mountain glaciation

The Fan Mountain glaciation is here named for glacial deposits found at the thresholds of cirques in the vicinity of Fan Mountain, 13 miles south of Shainin Lake (Fig. 3). Moraines of the Fan Mountain glaciation appear to be present in many cirques in the higher parts of the Brooks Range, generally above an altitude of 4,600 feet in the vicinity of Alapah Mountain (Fig. 10). The distribution of the deposits is shown in Fig. 3.

Most till of this glaciation is fresh and irregular. Tundra vegetation has not yet advanced over these young deposits, which are composed of unsorted, locally derived rock fragments. They do not differ from moraines now forming at the front of existing glaciers. Precise differentiation between till ridges that are being formed at present and Fan Mountain moraines is difficult near the snouts of a few of the larger glaciers where the ice may abut against the Fan Mountain moraines or may actually have overridden them. Kettle lakes are present behind a few of the Fan Mountain moraines. In some ice-free cirques the deposits form an irregular mantle of till; in others they form a single terminal moraine, and in still others as many as six closely spaced morainal ridges. Drainage in these essentially undissected morainal areas is largely by subsurface seepage.

Small glaciers confined to the higher, larger, and mainly north-facing cirques characterized the Fan Mountain glaciation. Numerous ice-free cirques with Fan Mountain moraines indicate the disappearance of many of these glaciers. This glaciation appears to represent an extremely recent and short-lived advance.

Photo: U. S. Navy

Fig. 10. Deposits of the Fan Mountain glaciation in cirques at the head of Kayak Creek, 3 miles northwest of Fan Mountain. The floors of the cirques are approximately 5,000 feet above sea-level; the peaks rise to nearly 7,000 feet.

Loess deposits

Deposits formed primarily by wind action, but possibly including some material of fluvial origin, overlie the terrace deposits along the Colville River (Fig. 11) and the lower Chandler, Anaktuvuk, and Nanushuk rivers. Similar deposits are probably present elsewhere, but they have not been recognized. The loess deposits are from 15 to 30 feet thick, buff to brown in colour and are composed of silt particles. The areas covered by loess are of uniformly low relief, have only a few, poorly integrated streams and contain numerous large thermokarst lakes.

Correlation of the glacial sequence

The glacial deposits in northern Alaska have not been dated by carbon-14 analyses. Correlation with other glaciations in Alaska and with the standard North American stages is by comparison of physical characters produced by mass-wasting and erosion. The vegetation of the Arctic Slope of the Brooks Range is predominantly tundra, with willow and alder groves along the larger rivers. Surficial modification of glacial deposits by mass-wasting would therefore be considerably different from that in forested areas.

Each of the six glacial advances recognized in the northern Brooks Range and on the Arctic Slope produced deposits with different characteristics. The differences between some deposits are more pronounced than those between others. They are used together with the patterns of distribution in establishing the sequence shown in Table 1. The most striking

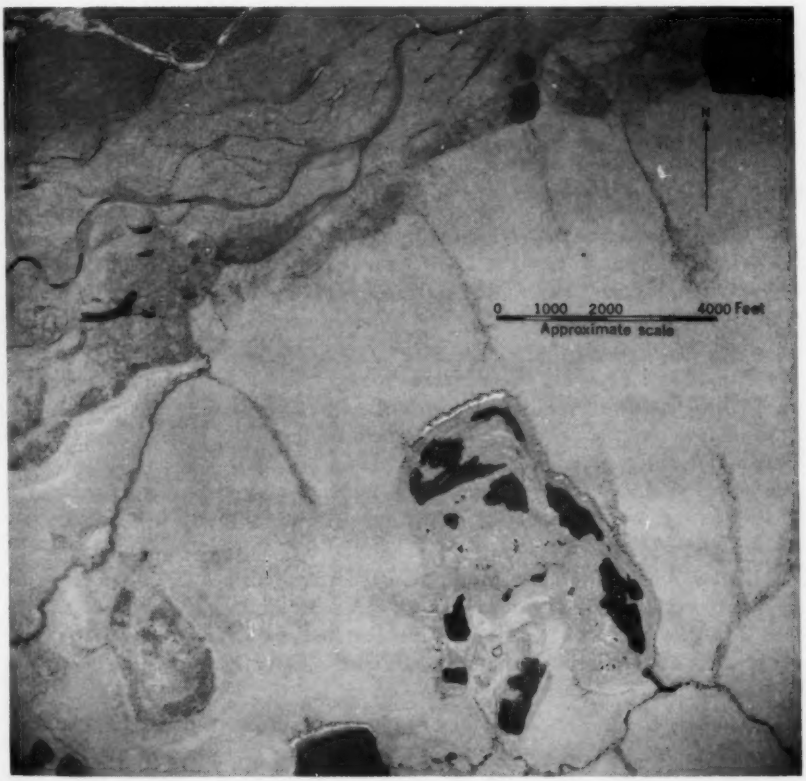

Photo: U. S. Navy

Fig. 11. Thermokarst lakes in loess overlying terrace gravel in the loess belt along the south side of the Colville River near Umiat.

feature of the Anaktuvuk and Sagavanirktok deposits is their apparently very great age. They are clearly much older than 8,000 years, the age that may be provisionally assigned to the Echooka. In the absence of means of dating other than by physical characteristics, the authors believe that a precise age assignment, other than pre-Wisconsin, would be ill-advised.

Itkillik and Echooka deposits are relatively fresh and somewhat similar in appearance, although minor physical differences and different distributional patterns make a distinction between the two glaciations possible. The possibility that the Echooka represents a major readvance or a period of stand-still within the Itkillik glaciation cannot be ruled out definitely at the present time. The fresh appearance of the drift of these two glaciations would seem to indicate that they are no older than early Wisconsin.

Table 1. Tentative chronology of the glacial sequence in a part of the Arctic Slope of the Brooks Range, Alaska.

Recent	Fan Mountain
Late Wisconsin	Alapah Mountain
Early Wisconsin	Echooka Itkillik
Pre-Wisconsin	Sagavanirktok Anaktuvuk

Three samples of peat obtained by Boston University Physical Research Laboratories from a core in loess near Umiat have been dated by the Lamont Geological Observatory (Broecker, Kulp, and Tucek, 1956). The oldest sample was dated as 8,300 ± 270 years before the present. The other two samples have dates of 3,530 ± 150 years and 5,890 ± 170 years.

The oldest sample, taken from pollen core I of Livingstone (1957, pp. 257-259), was correlated by him with a similar zone at Chandler Lake. Because the precise relationship of glacial deposits to pollen zones in the Chandler Lake region has not been worked out, it seems best to consider this date of slightly more than 8,000 years as representative only of a minimum age for the Echooka glaciation. It may actually be considerably older.

Deposits of the Alapah Mountain glaciation are only slightly modified by weathering. They are, however, covered by tundra, which would indicate that they were not formed within the last one or two centuries. Consequently, the Alapah Mountain advance is tentatively dated as late Wisconsin.

Deposits of the Fan Mountain glaciation, composed of completely bare rock debris are probably of Recent age.

The chronology suggested in Table 1 is somewhat different from that suggested by Péwé *et al.* (1953, p. 13) or Karlstrom (1957, p. 74). Karlstrom changed the original correlation of Alaskan glacial sequences so that advances older than Naptowne, Brooks Lake, and Farewell were correlated

with the Illinoian, Kansan, and Nebraskan stages. The greatest change in physical characteristics of the deposits produced by mass-wasting, as described in Péwé *et al.*, appears between Wisconsin and pre-Wisconsin deposits. In the area described in this report it occurs between the deposits of the Sagavanirktok and Itkillik advances; therefore, the Itkillik advance has been dated as early Wisconsin, as originally stated in Péwé *et al.* (1953).

Quinn (1957, p. 164) has discussed the possible contemporaneity of Texas and Alaska Pleistocene alluviations. Two distinctly different sets of terraces are recognized. Deposits of the older terraces are considered to be pre-Wisconsin and those of the younger terraces are thought to be Wisconsin. Quinn attributes differences in distribution and physical character of these terraces to major climatic changes during the Pleistocene. Differentiation of glacial deposits on the Arctic Slope of the Brooks Range into older (pre-Itkillik) and younger (Itkillik and post-Itkillik) glaciations may be a reflection of similar Pleistocene climatic changes.

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Award of Institute research grants

The following have been awarded research grants for field and laboratory investigations by the Institute for 1958 from the Sir Frederick Banting Fund:

DAVIS, IAIN. McGill University, Montreal, Quebec, Canada.
Permafrost studies on Southampton Island, N.W.T.

DERBYSHIRE, EDWARD. McGill University, Montreal, Quebec, Canada.
Extension of previous field work to include the west side of the main watershed of New Quebec-Labrador and the critical area of the Howells Valley, in an attempt to elucidate the nature of the late Wisconsin deglaciation of an important and unknown area.

DRIVER, PETER MICHAEL. McGill University, Montreal, Quebec, Canada.
Investigation of breeding behaviour and ecology of diving ducks at Ungava Bay.

GAGNE, RAYMOND CLOVIS. University of Montreal, Montreal, Quebec, Canada.
To make a phonemic analysis of an Eastern Arctic Eskimo dialect as a necessary first step leading to the establishment of a standard Canadian Eskimo orthography.

KROGER, ANNEMARIE. McGill University, Montreal, Quebec, Canada.
Study of the glacial morphology of the area north of Dease Arm, Great Bear Lake, between Dease and Haldane rivers.

MACINNES, CHARLES DONALD. Bishop's University, Lennoxville, Quebec, Canada.
Preliminary survey for an oceanographic study in Ungava Bay.

MACINTYRE, ROBERT JOHN. McGill University, Montreal, Quebec, Canada.
An intensive study of the biology of marine species of the amphipod *Gammarus* in Ungava Bay.

MACPHAIL, JOHN DONALD. University of British Columbia, Vancouver, B.C., Canada.
Collections for distributional and taxonomic studies of fishes in southeastern Alaska and southwestern Yukon Territory. A zoogeographic study of the fish fauna of the upper Yukon River and the North Pacific coastal drainages.

MACPHERSON, ELIZABETH. Carleton College, Ottawa, Ontario, Canada.
Collections of blood slides and examination of them for blood parasites. Collection of other endoparasites, fish otoliths, and plants, in Prince of Wales Island, N.W.T.

MYRES, MILES TIMOTHY. University of British Columbia, Vancouver, B.C., Canada.
To describe and film sexual and non-sexual display behaviour of at least two species of eider ducks at Homer and Point Barrow, Alaska, as part of a study of the ethology of sea ducks, in which courtship displays are being used as indicators of the systematic relationships of genera and species.

TOMLINSON, ROGER FRANK. McGill Subarctic Research Station, Knob Lake, Quebec, Canada.
Glaciological and geomorphological studies in the Kaumajit Mountains of northeastern Labrador.

WEEDEN, ROBERT BARTON. University of British Columbia, Vancouver, B.C., Canada.
To continue studies begun in 1956 on the ecological factors determining the distribution of ptarmigan in western North America.

NORTHERN NEWS

Salmon investigations on the Koksoak River, 1957

Continuing the investigations that were begun at the George River in 1956, the salmon population of the Koksoak River was examined during the summer of 1957. Sampling was by the use of nylon gill nets placed at various stations along the river and its two major tributaries, the Kaniapiskau and the Larch. A fairly large sample of juvenile salmon was obtained by angling. It was found that probably the majority of the Koksoak salmon travel up the Larch River to spawn. How far they can ascend is not known, but they do so for a minimum of 60 miles and probably can go much farther. Only the lower 20 miles of the Kaniapiskau River are accessible to salmon, beyond this they are stopped by the 60-foot Limestone Falls. On the basis of catches of adult salmon twenty times as many ascend the Larch than the Kaniapiskau; however, since only one salmon was taken in the Kaniapiskau in 30 net nights, this estimate is probably not reliable.

The Eskimo salmon fishery at Fort Chimo was observed when it was in full swing and a good sample of the catch was examined. This fishery can hardly be described as intensive and undoubtedly the Eskimo could obtain far more salmon. The major obstacles are (1) the high cost of gear; (2) the difficulties of fishing the Koksoak River at Fort Chimo

due to the large size of the stream and the strong tidal currents; (3) the coincidence of the fishing season and the time of seasonal employment in loading and unloading the ships. This occupation is very attractive because the Eskimo can earn $1.40 per hour, whereas fishing does not bring in cash, only a supply of dog food for the winter.

Since the supply of dog food runs out at Fort Chimo usually around Christmas and the dogs from then on have to be given imported food, something ought to be done to improve this situation. Perhaps a co-operative scheme might be organized whereby a few men could catch sufficient fish to supply the community for the whole winter and sell them to those who are employed in other ways.

In addition to collecting, physiological studies on the speckled trout and the Atlantic salmon were continued. These involved measurements of the basal oxygen consumption of these fish under different temperature regimes at different seasons. Results obtained for the speckled trout in previous years suggested that diurnal temperature fluctuations at low temperatures had an exaggerated effect on the oxygen consumption of the fish. To complement the hourly measurements of temperatures made in conjunction with the physiological experiments, a number of continuous recordings were obtained using a thermograph kindly supplied for this purpose by the Taylor Instrument Company.

The results of these investigations, which were supported by a Banting Fund grant from the Arctic Institute of North America and a grant from the Department of Fisheries, Quebec, will be published in detail later.

G. Power

GEOGRAPHICAL NAMES IN THE CANADIAN NORTH

The Canadian Board on Geographical Names has adopted the following names and name changes for official use in the Northwest Territories and Yukon Territory. For convenience of reference the names are listed according to the maps on which they appear. The latitudes and longitudes given are approximate only.

Lubbock, 105 C/4			
(Adopted March 7, 1957)			
Hawdt Lake	60°07′N	133°35′W	
Nokudsay Lake	60°08′	133°37′	
(Adopted May 2, 1957)			
Name confirmation			
Haunka Creek	60°15′N	133°44′W	not Stony Creek
Fort Norman, 96C			
(Adopted March 7, 1957)			
McConnell Range	64°20′N	124°10′W	
Red Dog Mountain	65°12′	125°39′	
Red Dog Creek	64°11′	125°40′	
Moose Nest Creek	64°12′	125°48′	
Moose Prairie	64°03′	125°58′	
Middle Creek	64°15′	125°09′	
Little Smith Creek	64°26′	124°41′	
Big Smith Creek	64°35′	124°53′	
Little Birch River	64°41′	125°06′	
Seagull Island	64°45′	125°05′	
Fall Stone Lake	64°43′	125°42′	
Yellow Lake	64°39′	125°39′	
Four Mile Island	64°54′	125°31′	
Lake Tagatui	64°58′	125°12′	
Police Island	64°51′	125°10′	
MacKay Range	64°43′	125°43′	not MacKay Mountains

(Adopted August 1, 1957)
Name confirmations

Burnt Point	64°16′N	124°51′W	
Loon Lake	64°57′	125°03′	
Jackfish Lake	64°48′	125°18′	

Braeburn Lake, 105 E/5
(Adopted March 7, 1957)

Contact Mountain	61°27′N	135°59′W	
Belleview Mountain	61°21′	135°53′	
Corduroy Mountain	61°19′	135°55′	
Hull Mountain	61°17′	135°57′	
Kingston Mountain	61°16′	135°54′	
Ottawa Mountain	61°17′	135°50′	
Fox Lake	61°17′	135°31′	not Richthofen Lake nor Richthofen (Fox) Lake

(Adopted April 4, 1957)

Big Hill	61°25′N	135°34′W	
The Dome (hill)	61°24′	135°31′	not Dome (hill)

Mount Bryde, 105 C/3
(Adopted March 22, 1957)

Spawn Lake	60°03′N	133°24′W	
Pike Lake	60°08′	133°22′	

(Adopted May 2, 1957)

Naystayo Lake	60°05′N	133°22′W	not Anderson Lake
Nuska Lake	60°06′	133°23′	not Wolverine Lake

King William Island, 67 SW and 67 SE
(Adopted March 7, 1957)

Malerualik Lake	68°34′N	97°19′W	
Douglas River	68°35′	97°08′	
Longfellow Inlet	68°07′	98°30′	
De Haven Island	68°04′	98°34′	
Kettle Cove	68°04′	98°34′	
Knud Inlet	68°22′	97°39′	
Padliak Inlet	69°26′	103°10′	
M'Clintock Bay	68°39′	97°43′	not Gilder Bay
Koka Lake	68°31′	96°22′	not Koaka Lake nor Kokak Lake nor Kookar Lake nor Kaa-aak-ta Lake nor Kaorka Lake nor Qoqa Lake

Ogden Bay, 66 NW and 66 NE
(Adopted April 4, 1957)

Sherman Basin	67°47′N	97°35′W	
Ermine Harbour	67°54′	98°13′	
Falcon Inlet	67°45′	98°04′	
School of Whales (islets)	67°48′	98°04′	
Schwatka Islands	67°51′	97°50′	
Crescent Harbour	67°52′	98°08′	
Altered application			
Sherman Inlet	68°00′	98°21′	

Rae Strait, 57 SW and 57 SE
(Adopted April 4, 1957)

Gibson Peninsula	68°46′N	95°25′W	
Neumayer Peninsula	68°39′	95°51′	
Astrup Island	68°39′	95°30′	
Rasmussen Basin	68°26′	95°45′	
Petersen Bay	68°37′	95°57′	not Peterson Bay nor Peterson's Bay nor Petersen's Bay
Betzold Point	68°38′	95°51′	not Bezold Point
Lund Islet	68°36′	95°52′	not Anton Lund Islet

Camsell River, 86 SW and 86 SE
(Adopted May 2, 1957)

Thomas Lake	65°08′N	119°10′W	
Farrell Lake	65°01′	119°34′	

Cambridge Bay, 77 SW and 77 SE
(Adopted May 2, 1957)
Name change

Augustus Hills	69°08′N	105°27′W	not Augusta Hills

Bathurst Inlet, 76 NW and 76 NE
(Adopted May 2, 1957)
Name changes

Amagok Creek	66°34′N	107°37′W	not Wolf Creek
Hiukitak River	67°08′	107°15′	not Hiokitak River

Chart 5149, Hudson Bay, northern portion
(Adopted May 2, 1957)

Jigging Point	64°14′N	94°52′W	
M'Clure Point	64°04′	81°17′	
Walrus Island	63°56′	89°34′	
Pricket Point	63°28′	76°33′	
Minion Rock	63°23′	76°33′	
Mirage Islands	62°39′	91°48′	
Prairie Bay	62°50′	92°05′	
Cairn Cove	62°50′	81°56′	
Calanus Bay	62°55′	83°00′	
Poillon Point	63°56′	89°00′	not Poillion Point nor Pullen Point
Morso Islands	62°02′	93°42′	not Morso Island
Magnet Island	60°16′	77°38′	not Magnet Islands
Name confirmations			
Enukso Point	64°34′	78°10′	not Inuksuk Point
Igloo Point	62°21′	92°06′	not Iglu Point
Pangertot Peninsula	62°36′	92°20′	not Pangnektok Peninsula
Kingnait Range	64°30′	77°00′	not Kingait Range
Altered applications			
Whale Point	64°12′	88°00′	
Yellow Bluff	64°22′	87°50′	
Gore Point	64°05′	81°25′	
Hut Point	63°18′	84°32′	
Prairie Point	63°59′	83°10′	
Waters Island	59°03′	80°44′	
Two Brothers (islands)	58°40′	79°55′	
Thompson Harbour	60°14′	77°38′	
Deletion			
Back Island	63°44′	80°07′	

Minto Lake, 115 P/9
(Adopted May 2, 1957)

Turnip Creek	63°42′N	136°16′W
Carlson Creek	63°41′	136°13′
McRae Creek	63°45′	136°12′
Dredge Creek	63°45′	136°11′
McIntyre Creek	63°43′	136°06′
Jarvis Creek	63°42′	136°07′
Sunday Creek	63°41′	136°10′
Paradise Creek	63°42′	136°07′
(Adopted July 4, 1957)		
Additional names		
Liberal Ridge	63°37′N	136°08′W
Conservative Ridge	63°37′	136°25′
Conservative Trail	63°42′	136°15′

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